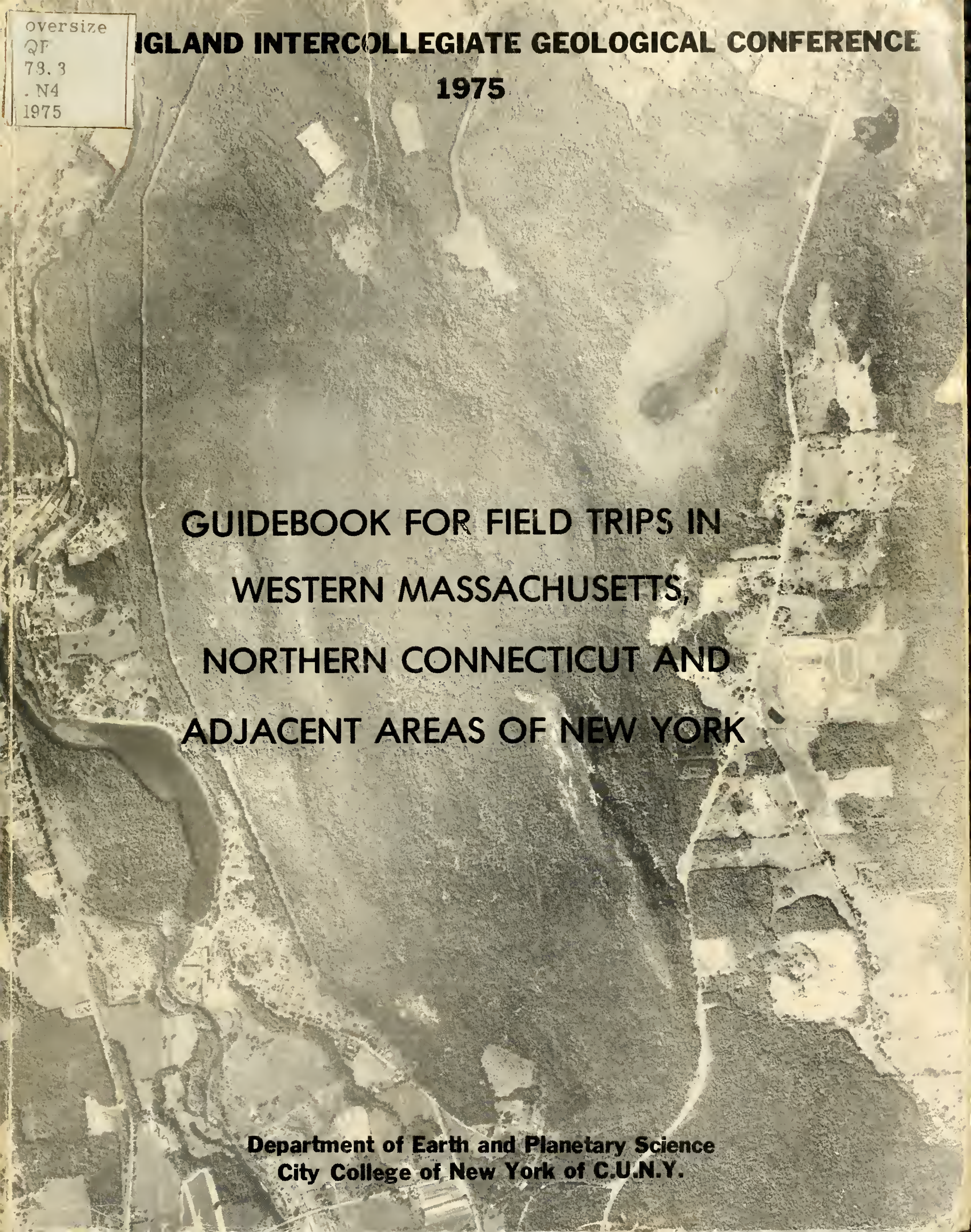

IGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE

1975

GUIDEBOOK FOR FIELD TRIPS IN WESTERN MASSACHUSETTS, NORTHERN CONNECTICUT AND ADJACENT AREAS OF NEW YORK

Department of Earth and Planetary Science
City College of New York of C.U.N.Y.

NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE

67TH ANNUAL MEETING

GUIDEBOOK FOR FIELD TRIPS IN WESTERN MASSACHUSETTS, NORTHERN CONNECTICUT AND ADJACENT AREAS OF NEW YORK

October 10, 11, and 12, 1975

Nicholas M. Ratcliffe, Editor
City College of C.U.N.Y.
New York, New York

A limited number of guidebooks will be available for purchase after the meeting for $6.00 (includes postage and handling). Order from:

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FOREWORD

This year's NEIGC has consciously been restricted in focus and areal coverage in the hope of restoring some of the collegial nature of the conference. Trips will be fewer in number and will be run from a central meeting point. Trip leaders kindly consented to repeat trips so that participants may have the maximum flexibility in choosing trips. Credit for the success of the meeting rests almost entirely with the authors. I did little except collect their products, if sometimes rather belatedly. I extend my sincere thanks to all the trip leaders for their very considerable efforts.

Much of the information presented in this guidebook stems from mapping programs supported over the past ten years by the U. S. Geological Survey in Cooperation with the Massachusetts Department of Public Works and the Connecticut Natural History and Science Survey. Special thanks are due to the U.S.G.S. for its help in processing manuscripts and for allowing its personnel to participate in this meeting.

Nicholas M. Ratcliffe
Editor

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Sunday trips

SOME BASEMENT ROCKS FROM BEAR MOUNTAIN TO THE HOUSATONIC HIGHLANDS*

By Leo M. Hall,[1] Henry L. Helenek,[2] Richard A. Jackson,[1]

Katherine G. Caldwell,[1] Douglas Mose,[3] and Daniel P. Murray,[4]

INTRODUCTION

This field trip is designed to illustrate some of the rock types and structural features present in Precambrian terranes between Bear Mountain, New York and the Housatonic Highlands in western Connecticut (Fig. 1). As indicated on Figure 1, exposures in the Hudson Highlands (Reading Prong), New Milford Massif, Bear Hill Massif, and the Housatonic Highlands will be visited. Basal Paleozoic rocks, the Poughquag Quartzite and Lowerre Quartzite, in contact with Precambrian gneisses are present at Stops 3, 4, 5, and 6 (Fig. 1) and it is interesting to compare the eastern more feldspathic and per-aluminous Lowerre facies to the western clean quartzite and arkosic Poughquag facies.

The Hudson Highlands are conveniently divided into two sections, the Western Highlands and the Eastern Highlands, by the Ramapo-Canopus fault zone (Fig, 1). This fault zone has had a long complicated tectonic history (Ratcliffe, 1971) and the basement rocks on opposite sides of this zone are dissimilar to a certain extent. The most abundant rocks in the Western Highlands are hornblende granite and other granitic gneisses along with charnockitic gneisses (Stops 1 and 3) whereas biotite granitic gneiss and amphibolite, both of uncertain origin, and biotite-hornblende-quartz-feldspar gneiss (Stop 2) are the most abundant rocks in the Eastern Highlands. Paleozoic deformation and metamorphism has had a much greater effect on rocks in the Eastern Highlands. Aeromagnetic studies (Harwood and Zietz, 1974) indicate a distinct difference in the magnetic character of the rocks in each region, with those in the Eastern Highlands having a much lower magnetic intensity than those in the Western Highlands. Rocks in the Eastern Highlands have magnetic signatures similar to the Fordham Gneiss to the south and to the basement massifs in western Connecticut and thus seem to be more akin to them than to the Western Highlands (Fig. 1). Lithic similarities also exist between the Fordham Gneiss and some of the rocks in the Eastern Highlands and it appears that they are at least parially equivalent stratigraphically and very similar structurally. A definitive understanding of the relationship between them is a major problem that needs detailed study.

Paleozoic rocks in the region display an increasing grade of metamorphism toward the east, up to the sillimanite-orthoclase zone (Balk, 1936; Vidale, 1974), and apparently the more intimate structural involvement of the basement and cover rocks toward the east is directly related to this increase in metamorphic grade. This intimate involvement between basement and cover is evident in the Manhattan Prong toward the south (Hall, 1968) and will be observed in western Connecticut at Stops 4 and 5 on this field trip (Fig. 1). The exposure of Precambrian gneiss resting directly on top of the Cambrian Lowerre Quartzite at Stop 4 along

*Helenek, Murray, and Mose provided the data for the first three stops of the field trip and Jackson, Caldwell, and Hall are responsible for the last three stops. The text for the guidebook has been jointly prepared by all participating leaders.

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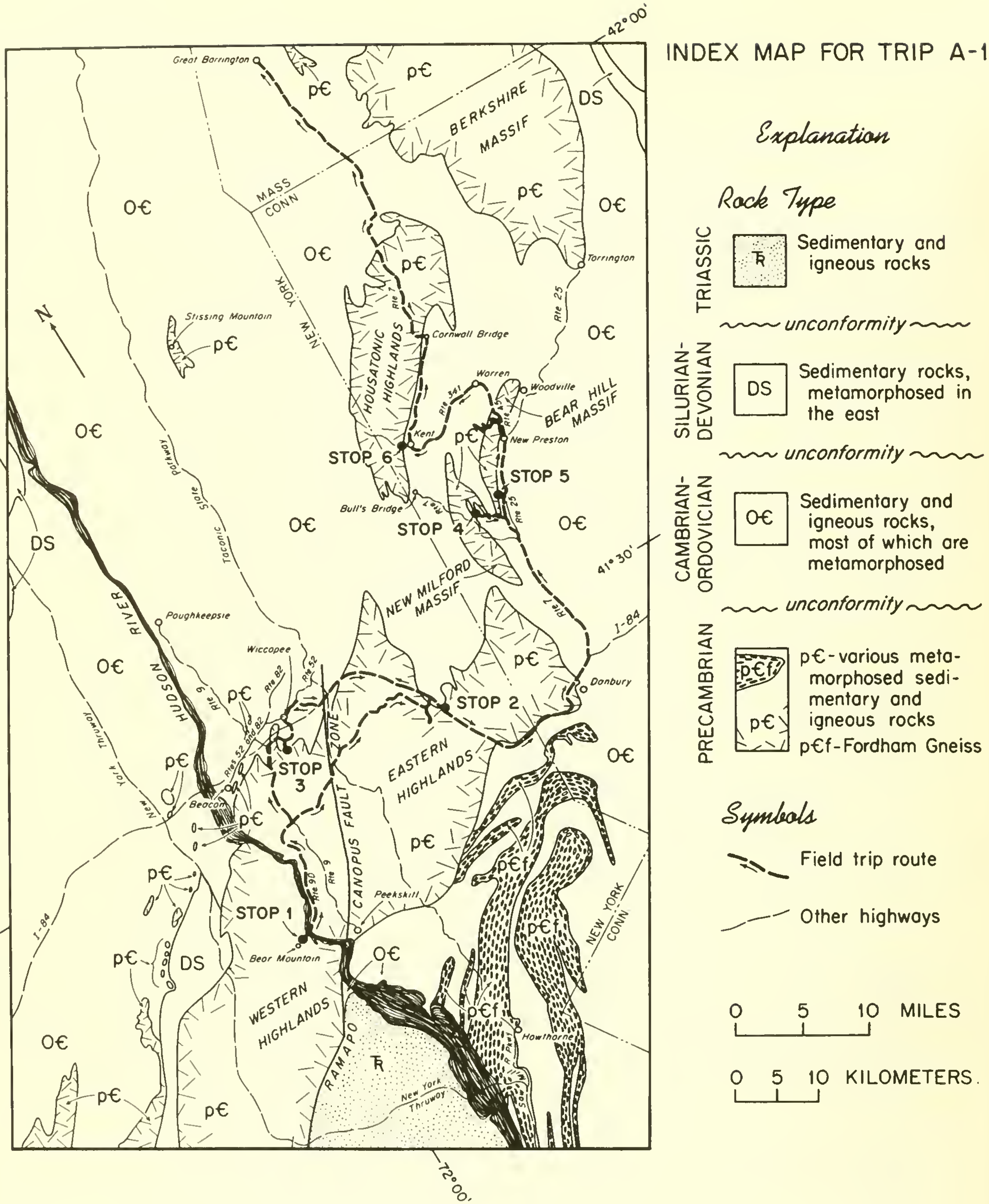


Figure 1. Index map and route for field trip A-1.

the east side of the New Milford massif is particularly spectacular and betrays the complex structural character of the New Milford massif which is currently being studied by Katherine Caldwell. The Bear Hill massif (Fig. 1) was also intimately involved with the deformation of the cover rocks and is currently being studied in detail by Richard Jackson. Gregory and Robinson (1907) and Agar (1929) are among the earlier geologists who studied the rocks in western Connecticut and interpreted the gneisses in the Bear Hill region to be Precambrian. Subsequent work by Gates (1952) led to the interpretation that these gneisses are part of the Paleozoic section. Jackson has found the Cambrian Lowerre Quartzite unconformably truncating rock units in the gneisses (Stop 5) and consequently has reinstated the earlier interpretation that these gneisses are Precambrian basement. The bedrock geology of the New Milford and Bear Hill massifs and their associated cover rocks is strikingly similar to that found in most of the Manhattan Prong and may also be similar to that associated with the Eastern Highlands. This is true with respect to the nature of the cover rock stratigraphy, some of the basement rock types, and the intimate structural relations between basement and cover. The pink quartz-feldspar gneiss and the pink augen gneiss (Stops 5 and 6) in the basement in western Connecticut bear a striking resemblance to the Yonkers Gneiss in the Manhattan Prong.

The autochthonous as opposed to allochthonous nature of the basement rocks that will be visited on this trip remains as a regional geologic problem. Various interpretations of these relationships have been suggested (Isachsen, 1964) and Dallmeyer (1974). This specific problem will not be addressed directly on this field trip however most people would agree that if any large scale thrusting of the basement has occurred it must also have involved at least part of the cover rock section. On the other hand, the shearing along the Poughquag-Precambrian contact at Stop 6 (Fig. 1) suggests the possibility that the cover may have broken loose and sheared over the basement at least locally. Present and future studies will undoubtedly shed more light on this regional problem.

ACKNOWLEDGEMENTS

We wish to thank Marie Litterer for drafting the figures in exceptionally fine fashion. We are also appreciative of the cooperation extended to us by the personnel of the Harriman State Park, the Sharpe Environmental Center, the New York State Police, the Eliot D. Pratt Education Center, and to Mr. and Mrs. Lardner for allowing us to park on their property at Stop 4.

WESTERN HIGHLANDS

Lithology - The most common rock types and their approximate relative abundances are hornblende granite and granitic gneiss (35%), various charnockitic quartz-plagioclase gneisses (27%), amphibolite and related rocks (15%), paragneiss (13%), and alaskite (5%). Biotite granite gneiss (3%) and calcareous, ferruginous and quartzitic metasediments (2%) are subordinate. Most of these rocks are interpreted as either meta-volcanic or meta-sedimentary on the basis of bulk composition and association. Zircon populations of the hornblende granitic rocks are characteristic of plutonic rocks (Eckelmann and Helenek, 1975) thus the hornblende granites probably represent sills and phacoliths intruded prior to and during the major deformation and metamorphism. Alaskite, Canada Hill granite and the Canopus pluton (Ratcliffe et.al., 1972) are the only unequivocal Precambrian intrusive igneous rocks.

Stratigraphy - The following stratigraphic sequence has been identified in the Western Highlands (Helenek, 1971; Jaffe and Jaffe, 1973):

Unit C. Migmatitic paragneiss and Canada Hill type granite with subordinate amphibolite and quartz-plagioclase leucogneiss and minor calcareous ferruginous and quartzitic metasediments.

Unit B. Quartz-plagioclase gneisses with subordinate amphibolite and calc-silicate gneiss and minor additional meta-sediments.

Unit A. Hornblende granite and granitic gneiss.

The history represented by this stratigraphy has been interpreted in various ways. Offield (1967) proposed that Unit B was metamorphosed and intruded by Unit A producing an ancient Precambrian basement which was subsequently eroded. Unit C was then deposited unconformably on this ancient basement and the entire section was then metamorphosed and deformed. Helenek (1971) concluded that Unit A represents remobilized basement that was initially overlain by Units B and C and that the Canada Hill type granite was derived by partial melting of paragneiss. Jaffe and Jaffe (1973) interpret Unit C to be the oldest in the sequence and Unit A to be a thick section of meta-rhyolite at the top.

Structure - At least three periods of folding have effected the rocks in the Western Highlands and the earliest fold generation (F_1) resulted in regional isoclinal folds. The second generation of folds (F_2) developed as plane cylindrical, isoclinal folds with southeast dipping axial planes and axes that plunge approximately N35°E at 10° parallel to the regional lineation (B_2'). A localized set of open folds (F_3) with nearly vertical axial planes and axes that plunge N45°E at 30° refolded elements of F_1 and F_2-folds. Data for the dominant fabric elements S_o and B_2' are presented in (Figure 2 A + B).

Metamorphism - Critical minerals in the mafic rocks are clinopyroxene-orthopyroxene-hornblende-biotite-quartz-plagioclase and opaques and orthopyroxene-garnet-hornblende-quartz-plagioclase-microperthite in hornblende granitic rocks. These phases imply lower granulite facies metamorphism, and are compatible with the pressure-temperature conditions established in the Western Highlands (Fig. 3) by Dallmeyer and Dodd (1971). The horizontal bar on Figure 3 indicates the pressure and was obtained by applying a modified version of the cordierite-garnet geobarometer described by Hutcheoson et. al. (1974) to compositional data for these minerals. Retrograde metamorphic effects increase markedly from the Hudson River eastward to the Canopus fault.

Geochronology - The following Rb/Sr whole rock ages have been determined by Mose et. al. (1975):

Unit	Rock Type	Age $\pm$ 1σ	$(^{87}Sr/^{86}Sr)_o \pm 1\sigma$
A	Hornblende granitic gneiss at Crows Nest	1169 $\pm$ 14 m.y.	0.7055 $\pm$ 0.0006
B	Quartz-plagioclase gneiss	1115 $\pm$ 85 m.y.	0.7033 $\pm$ 0.0007
C	Paragneiss	1139 $\pm$ 10 m.y.	0.7067 $\pm$ 0.0004
A	Hornblende granite at Bear Mtn.	1086 $\pm$ 8 m.y.	0.7020 $\pm$ 0.0005
C	Partial melt from paragneiss (Canada Hill type granite)	914 $\pm$ 12 m.y.	0.7193 $\pm$ 0.0005

These age determinations are in good agreement with the 1170 m.y. zircon age from the "Canada Hill gneiss" (quartz-plagioclase gneiss?) and the 1060 m.y.

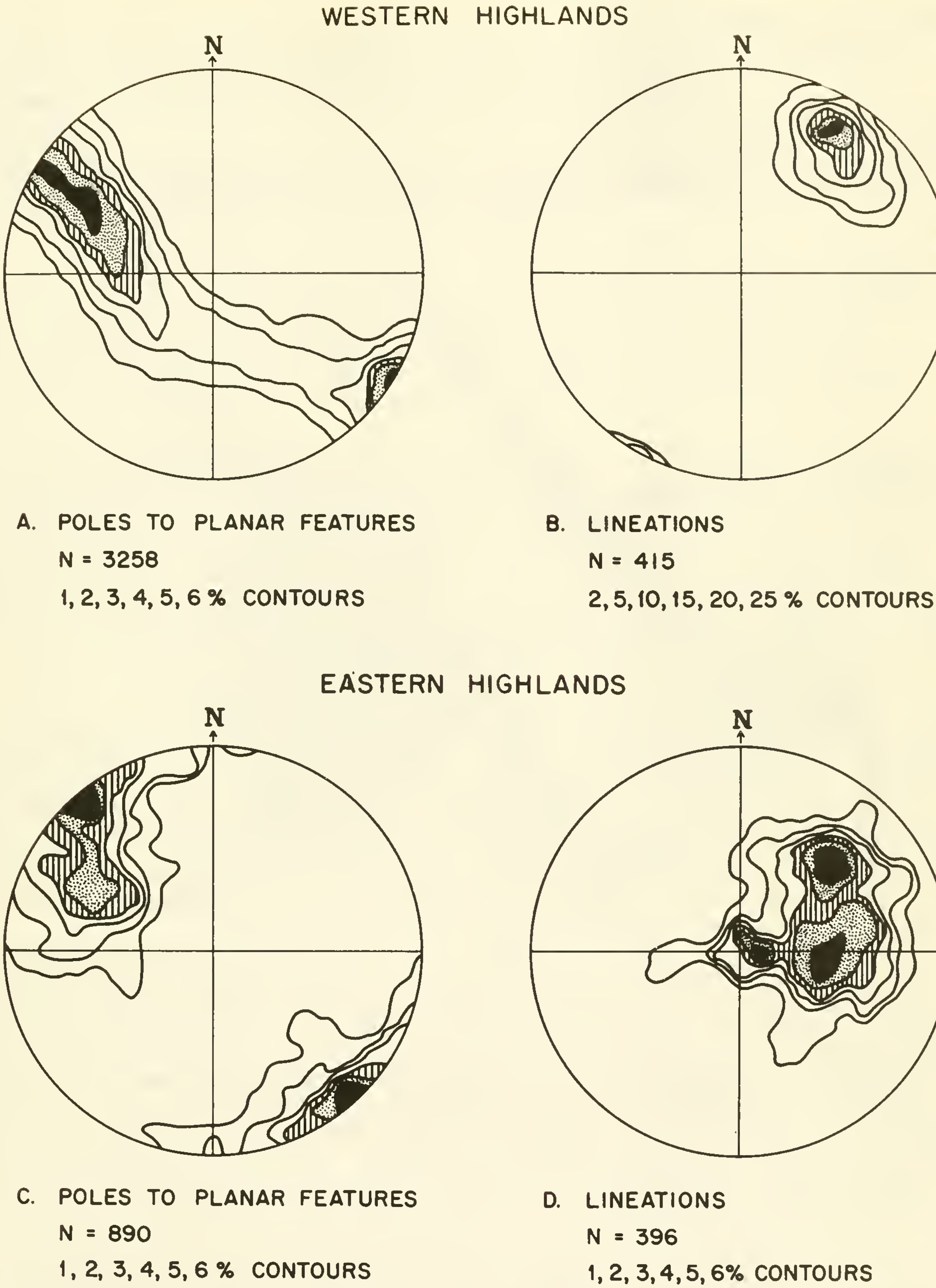


Figure 2. Equal-area diagrams summarizing planar and linear data for the Western Highlands (A and B) and the Eastern Highlands (C and D).

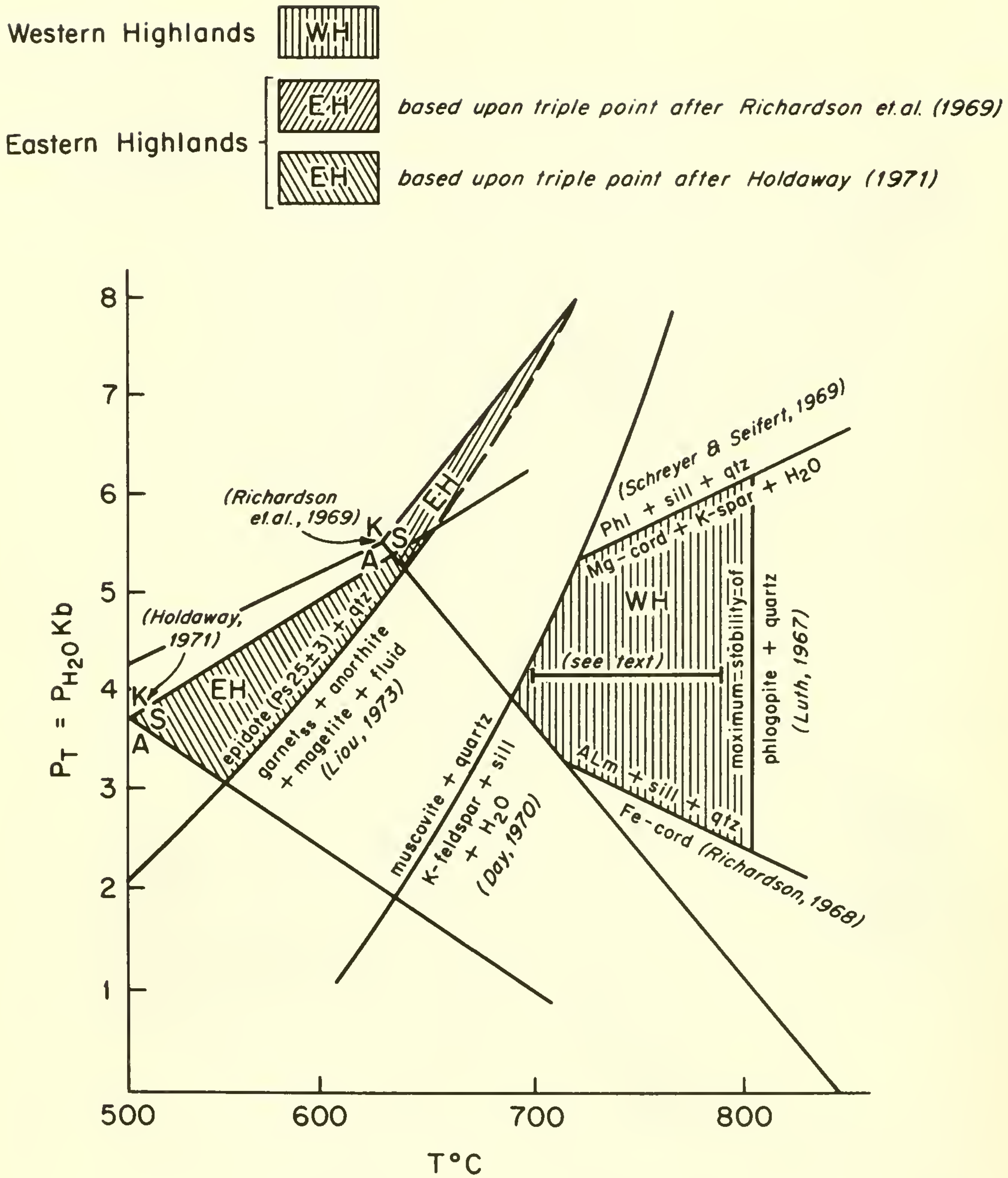


Figure 3. P-T conditions during metamorphism in the Hudson Highlands.

zircon age from hornblende granite at Bear Mountain (Tilton et al., 1960). K/A age determinations using biotite from rocks west of the Hudson River are 829 ± 34 m.y. (1σ, 12 ages) (Tilton et al., 1960; Long and Kulp, 1962).

Discussion

Stratigraphic analysis and radiometric dates indicate that the Western Highlands consist of rocks that originated through clastic sedimentation and intrusive and extrusive activity between 1200 and 1075 m.y. ago. All of these rocks were involved in all phases of Grenville deformation.

The Canada Hill-type granite (914 ± 12 m.y.) is the youngest intrusive rock associated with the Grenville event and its age indicates a youngest time-limit for Grenville activity in the Hudson Highlands. High initial $^{87}Sr/^{86}Sr$ ratios in the Canada Hill and its association with paragneisses indicate that it was derived by partial melting of paragneiss.

K/Ar age determinations on biotites average 829 ± 34 m.y. but the interpretation that this average age represents a thermal event is in doubt because no igneous rocks are known to have formed at this time. Another interpretation is that the biotite in these rocks reached its argon retention temperature of about 300°-350°C about 830 m.y. ago. Using biotite-hornblende pairs, with hornblende ages of about 900 m.y. and a hornblende retention temperature of about 500-550°C, Sutter and Dallmeyer (1972) proposed that the late Precambrian uplift occurred at a rate of about 10^{-5} meters per year.

EASTERN HIGHLANDS

Lithology - Preliminary work indicates that the relative abundance of the major rock types in the Eastern Highlands is biotite granite gneiss (58%), amphibolite (23%), and a layered biotite-hornblende-quartz-plagioclase gneiss (13%). Paragneiss (4%), quartz-plagioclase gneiss (1%), hornblende granitic gneiss (1%) and marble (trace amounts) are subordinate rock types. The hornblende granitic gneiss is an igneous rock and the biotite granite gneiss and amphibolite have uncertain origins whereas the remaining rocks are meta-sediments. Stratigraphic relationships have not been established in the Eastern Highlands.

Structure - The style and intensity of deformation of rocks in the Eastern Highlands differs from that in the rocks of the Western Highlands (Fig. 2) due mainly to penetrative deformation of Precambrian rocks during the Paleozoic. Tight isoclinal folds, tentatively assumed to have developed during a Precambrian deformation, are the earliest folds (F_1) recognized in the Eastern Highlands. The dominant planar structure is compositional layering (S_o) and the axial surfaces of F_1-folds are refolded into similar, reclined and recumbent folds (F_2). F_1 and F_2 fabric elements in turn have been folded by (F_3-folds) and all of these earlier structural features have been deformed by an F_4-warping. Since F_2- and F_3-folds are coaxial, they probably represent a single phase of continuous deformation tentatively believed to have developed during Taconic deformation. The dating of the F_4-folding in this scheme is uncertain.

Metamorphism - Isograds delineated by Balk (1936) immediately north of the Hudson Highlands can be extrapolated across the Hudson Highlands. Reconaissance studies of mineral assemblages from Precambrian rocks in the sillimanite

zone indicate upper amphibolite facies conditions. Textural features associated with minerals in these assemblages formed during the F_2 folding. Local relict textures that pre-date F_2 are present, as are retrograde effects that post-date F_2. The calc-silicate assemblage, epidote (Ps_{27})-grandite (And_{51}-$Gross_{38}$-$Other_{11}$)-augite (Wo_{49}-En_{18}-Fs_{33})-plagioclase (An_{30})-quartz-magnetite formed at the time of F_2-folding and defines a narrow range of T and fO_2 (10^{-18} bars). This assemblage coupled with that of sillimanite-garnet-biotite-quartz-feldspar gneiss nearby, defines a pressure-temperature regime for this area as shown in Figure 3.

Geochronology - Rb/Sr whole-rock studies on samples of biotite granite gneiss, from an outcrop in the Eastern Highlands near the Hudson River one-half mile north of Peekskill, indicates that this rock formed at 1256 ±7 m.y. (initial $^{87}Sr/^{86}Sr$ = 0.7021 ±0.0002). Another whole-rock study on samples of biotite granite gneiss thought to be the same, in the Lake Carmel 7 1/2' quadrangle, yielded an age of 1296 ±18 m.y. (initial $^{87}Sr/^{86}Sr$ = 0.7032 ±0.0003).

The 1250-1300 m.y. age of the biotite granite gneiss may represent its time of origin by sedimentary or igneous processes. Thus the biotite granite gneiss in the Eastern Highlands is considerably older than rocks in the Western Highlands. Although a Grenville event has not yet been identified in the Eastern Highlands, radiometric dating associated with the Fordham Gneiss (Grauert and Hall, 1973) indicates a relatively short Grenville event, over a 100-200 m.y. span, which ended about 980 m.y. ago and a younger event, 550 to 600 m.y. ago, probably Avalonian, involving the formation of the Yonkers Gneiss. Other dates on the Yonkers Gneiss (Long, 1969) and the Pound Ridge Gneiss (Mose and Hayes, 1975) indicate an Avalonian event in the Manhattan Prong about 580 to 600 m.y. ago.

K/Ar and Rb/Sr single mineral studies have been conducted on specimens from the Hudson Highlands, the Manhattan Prong, and Dutchess County (Long, 1962; Clark and Kulp, 1968). K/Ar age determinations on micas collected from Paleozoic rocks in the sillimanite zone are 387 ± 43 m.y. (1τ, 4 dates) and those from Highlands rocks in the same zone are 389 ± 44 m.y. (1τ, 4 dates). On the other hand, K/Ar mica ages from the garnet and biotite zones in the Manhattan Prong and Dutchess County average 406 ± 36 m.y. (1τ, 6 dates) whereas ages from rocks in these zones of Paleozoic metamorphism in the Eastern Highlands are 755 ± 50 m.y. (1τ, 5 dates).

The K/Ar age determinations on mica in rocks from the Hudson Highlands and the surrounding meta-sediments may be interpreted in various ways. One interpretation, based on the oldest ages from the low grade meta-sediments in the Manhattan Prong,would have a metamorphic event about 460-480 m.y. ago and then a second metamorphic event about 390 m.y. ago based on the average age determinations from high grade zones. Another interpretation is that the K/Ar age determinations on micas from progressively metamorphosed Paleozoic rocks in areas of lower grade metamorphism in the Manhattan Prong indicate a minimum age of 410 m.y. for a major metamorphic event. The younger K/Ar dates, about 390 m.y., from rocks in higher grade areas result from argon loss during uplift and cooling. The second interpretation is preferred because it is corroborated by regional interpretations involving the Ramapo-Canopus fault zone and associated metamorphic and intrusive events (Mose, et. al., in press).

SUMMARY

The Eastern and Western Hudson Highlands have pronounced differences in lithology, structural geology and geochronology. The Western Highlands are underlain by a metavolcanic and metasedimentary sequence that was intruded by syntectonic granitic plutons during the Grenville orogeny (1100-900 m.y.). Most of the rocks in the Eastern Highlands have uncertain origins, but there is little or no evidence that a phase of granitic plutonism occurred there during the Grenville. The data presented above indicate fundamental differences between these two terranes. There are numerous plausible interpretations of the geologic relations between them, but juxtaposition of the Eastern and Western Highlands, possibly two different crustal blocks, by motion along the Ramapo-Canopus fault zone is presently thought to be the most favorable interpretation.

ROAD LOG

Stop 1. Bear Mountain. The field trip participants will assemble at the Bear Mountain Inn parking lot at 9:30 a.m. Stop 1 is a short walk north of Bear Mountain Inn to the roadway north of Hessian Lake (Fig. 4). The exposures that will be studied at this locality are in Harriman State Park and regulations prohibit destruction of the environment in any way. NO HAMMERING, PLEASE!

We are in the axial region of an F_1-antiform (Fig. 4) and the rocks exposed here are typical of those in the Western Highlands (Fig. 1). The petrography of these gneisses is briefly described as follows:

Hornblende granite - The modal composition ranges as follows: quartz (30-40%), mesoperthite (50-70%); accessory biotite, opaques, apatite and zircon. At this locality it is poorly foliated and consists of slightly perthitic microcline, plagioclase,(up to 50% of the feldspar)and equal amounts of biotite and hornblende.

Amphibolite - The amphibolites consist of plagioclase (An_{35-40}; 27-40%), hornblende (37-62%), augitic clinopyroxene (9-18%) and biotite (1-3%) along with accessory opaques, apatite, and zircon. A prominent hornblende lineation is typically present in the amphibolite.

Biotite-pyroxene-plagioclase gneiss (rusty-weathering) - Plagioclase (sodic andesine, 36%), biotite (32%), clinopyroxene (29%) and opaques including some graphite (3%) with accessory apatite and zircon compose this rock.

Canada Hill type granite - This massive rock is composed of quartz, gray-mottled microcline-microperthite and white plagioclase feldspar (An_{24}). Garnet, sillimanite, biotite and graphite are also present in various amounts along with traces of muscovite, apatite, sphene, zircon and tourmaline.

Hornblende granite (p6hg) is in the core of the F_1-fold (Fig. 4) and is flanked successively by interlayered amphibolite and rusty pyroxenic gneiss (p6am) and paragneiss (p6m). The contact between p6hg and p6am is folded into isoclinal digitations and compositional layering (S_o) is completely transposed into axial surfaces of F_1-folds.

The gently northwest plunging F_1-folds are on the lower, southwest, limb of a reclined isoclinal antiform(F_2-fold, Fig. 4). The F_2 folds are locally similar in style and typically are more open than the F_1 isoclinal folds. A prominent hornblende lineation is parallel to the F_2 fold axes, B_2'. F_2-folds

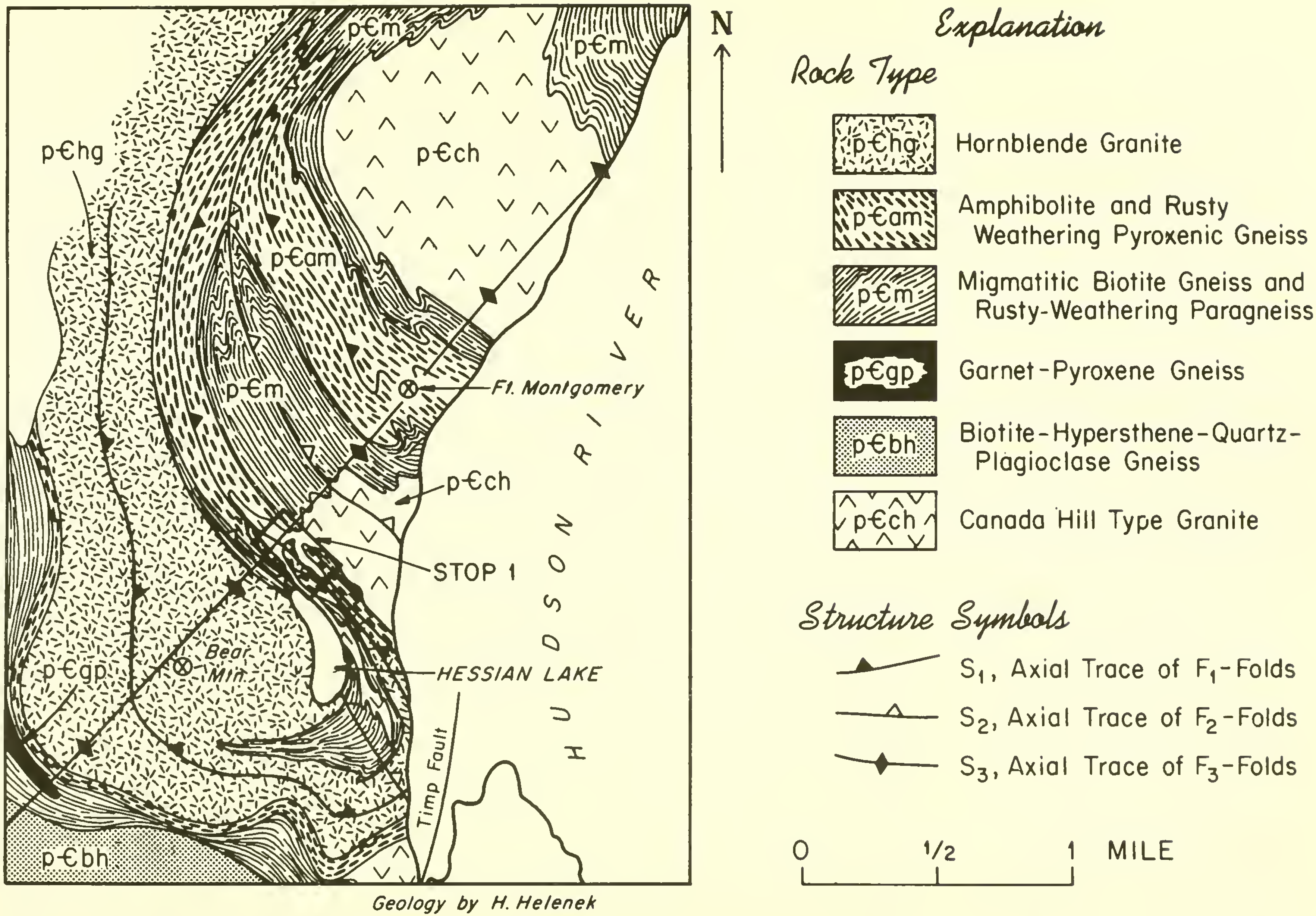


Figure 4. Geologic map of the region in the vicinity of Stop 1.

occur in the hornblende granite and, although not found in place at this locality, may be observed in a small boulder adjacent to the parking area. An F_2 axial plane foliation is present locally. Open F_3-folds have refolded fabric elements of the F_1- and F_2-folds. F_3-fold axes plunge 37°, N36°E and are almost coaxial with the F_2-fold axes.

Since the syntectonic hornblende granite has been involved in all phases of deformation, the Rb/Sr age of 1086 ± 8 m.y. and U/Pb zircon age of 1060 m.y. indicate the time of intrusion and the time of F_1-folding. This places an upper limit on Grenville diastrophism in this portion of the Western Highlands.

Field evidence in these exposures clearly indicates that Canada Hill type granite is a late tectonic intrusion which transects linear and planar features in all the gneisses. Thus the 914 ± 12 m.y. Rb/Sr whole rock age of the Canada Hill indicates a minimum age for Grenville diastrophism.

A subsidiary road log is included for an alternative short route in the event that we run short of time. The alternate route reverses the sequence of Stops 2 and 3 and appears in three segments: alternate route to Stop 3 at mileage 11.3, alternate route to Stop 2 at mileage 60.0, and alternate route to Stop 4 at mileage 35.7. Each segment of the subsidiary road log is separated from the main road log by lines. Those using this guidebook in the future should note that more regional geology will be seen on the main route.

Mileage Total	Mileage Interval	
0.0	-	Leave the Bear Mountain Inn in parking lot and turn left onto the service road. Proceed approximately two hundred yards and turn left onto Routes 9W and 202.
0.5	0.5	Start into the traffic circle and bear to the right and east across the Hudson River via the Bear Mountain Bridge. A view of Anthony's Nose is straight ahead at the east end of the bridge.
1.1	0.6	Turn left (north) at the east end of the bridge onto Route 9D and proceed toward Cold Spring.
9.0	3.4	Bear right onto the side road, Peekskill Rd., note the red Fire Department sign at the junction. Peekskill Rd. allows us to bypass the center of Cold Spring.
9.5	0.5	Turn right (northeast) onto Route 301 at the intersection of Peekskill Rd. with Main St. (Route 301) in Nelsonville.
11.3	1.8	Alternate route to Stop 3 from McKeel Corners follows:
0.0		Proceed north from McKeel Corners toward Fishkill by turning left onto Route 9 from Route 301.
4.7	0.7	Note the exposures of Precambrian at the east edge of the sand and gravel pit on the right (east) side of Route 9.
6.5	1.8	Turn right (east) onto Snook Rd. immediately before the I-84 route marker.
7.4	0.9	Snook Rd. joins Van Wyck Lake Rd. here. Proceed straight ahead on Van Wyck Lake Rd.

7.6	0.2	Junction with Cary Rd. is here. Bear to the right (southeast) staying on Van Wyck Lake Rd. Note the numerous exposures of quartz-plagioclase gneiss.
8.0	0.4	Sand and gravel pit on the right (southwest) side of the road.
8.3	0.3	Small bridge across a small stream. Approximately 600 feet southwest of this bridge there is a fine exposure of the basal Poughquag resting unconformably upon Precambrian migmatitic biotite gneiss. The easiest route to this south facing cliff is along the lane east of the stream.
8.4	0.1	Sharp curve in the road where it turns toward the east (left) and crosses the stream.
8.6	0.2	Turn right (south) into the entrance to Sharpe Reservation.

Follow the main road log from here beginning at mileage 53.4.

Continuation of Main Road Log

11.3	1.8	Intersection of Route 9 with Route 301 at McKeel Corners. Jog right and then left across Route 9 and proceed eastward on Route 301. Several exposures of Canada Hill type granite and paragneiss will be seen as you proceed.
13.7	2.4	Sampling locality for Rb/Sr whole rock dating (914 m.y.) of Canada Hill type granite is on the hill at the left.
14.7	1.0	Sampling locality for Rb/Sr whole-rock dating (914 m.y.) of the Canada Hill type granite.
15.3	0.6	Highly sheared Canada Hill type granite and paragneiss. The shearing here is probably associated with movement along the Ramapo-Canopus fault zone.
16.1	0.8	The Appalachian trail crosses the road (Route 301),here near the south end of Canopus Lake. The lake lies in the Ramapo-Canopus fault zone and as one looks northward along the lake the Western Highlands are toward the left and Eastern Highlands are toward the right. Exposures of grayish-white-weathering Canada Hill type granite are visible along the west shore of the lake.
20.6	3.1	Junction of Farmers Mills Rd. with Route 301 at sharp curve in Route 301. <u>Proceed straight ahead onto Farmers Mills Rd.</u> and continue to the junction with Route 52 north of Lake Carmel.
24.9	1.4	Exposure on the right is a sampling locality for Rb/Sr whole-rock dating (1296 $\pm$ 18 m.y.) of the Reservoir Granite.
25.5	0.6	Exposure on the right is a sampling locality for Rb/Sr whole-rock dating (1296 $\pm$ 18 m.y.) of the Reservoir Granite.
26.5	0.3	Bear right (south) onto Route 52 and proceed toward Lake Carmel.
28.5	0.4	Bad intersection at the junction of Routes 311 and 52. Turn left (east)onto Route 311 and proceed across the south end of Lake Carmel into the village and continue eastward on Route 311.

29.4	0.9	Turn onto the eastbound entrance ramp for I-84.
30.8	1.4	Large rock cuts on both sides of I-84 and the large cut on the westbound lane is Stop 2 (Figs. 1 and 5). Stop here to briefly observe the entire rock cut from a distance (refer to Figure 5).
32.8	2.0	Bear right onto the exit ramp for exit 19 which leads onto Route 312.
33.1	0.3	Turn left (northeast) onto Route 312 crossing the bridge over I-84 and then turn left onto the I-84 westbound entrance ramp and proceed west on the interstate highway.

35.7 2.6 Stop 2. I-84 Rock Cut. Pull vehicles well off the highway onto the shoulder. People are not permitted to cross the highway and it would be best if you stay in the grassy area off the shoulder after you leave your vehicle. This is definitely a camera stop!

The petrography of the Precambrian gneisses at this rock cut is summarized as follows:

Biotite granite gneiss - The dominant minerals are quartz, microcline/plagioclase (An_{15-30}) (ratio is varied) and biotite is minor. Accessory muscovite, zircon, epidote, apatite, and opaques commonly with rims of sphene are present.

Amphibolite and/or hornblende gneiss - Plagioclase (An_{20-50}, 35-40%), hornblende 40-45%), clinopyroxene (up to 12%) and biotite (up to 15%) are the common minerals. Hornblende rims are typically present on the clinopyroxene and accessories are quartz, apatite, calcite, sphene, garnet, epidote and opaques.

Biotite-hornblende-quartz-plagioclase gneiss - Quartz, plagioclase hornblende and biotite with minor microcline, epidote and sphene make up this rock.

Pyroxene-hornblende-quartz-plagioclase gneiss - plagioclase, quartz, hornblende and pyroxene along with minor garnet and sulfide constitute this rock.

Garnet-pyroxene-epidote nodules - Grandite (And_{51} $Gross_{38}$ $Other_{11}$) and ferroagite (Wo_{49} En_{18} Fs_{33}) with subordinate amounts of interstitial epidote, plagioclase and quartz in addition to accessory sphene, opaque, microcline and scapolite are present.

The earliest folds (F_1) are isoclinal and of varied size and orientation. A large F_1-fold extends through much of the outcrop (Fig. 5) and several smaller F_1-folds are present (Fig. 5, in the vicinity of point A). A pre-F_1 set of isoclinal folds may be present but their existence requires more thorough study to be proven.

Compositional layering (S_o) (Fig. 6A) and the axial surfaces of F_1-folds are refolded into reclined, recumbent, and inclined similar F_2-folds (Fig. 5, point B). Orientations of the axial surfaces (S_2') and fold axes (B_2') of the F_2-folds are shown on Figure 6 (C and D). Some of the more spectacular folds in the exposure are F_2-folds (Fig. 5, point C). Syntectonic granite accompanied the F_2-folding and is folded about B_2' (Fig. 5, point D). Late tectonic granite seams were injected preferentially along surfaces subparallel to the axial surfaces of F_2-folds (Fig. 5, point F) and a series of low angle thrust faults are also subparallel to the axial surfaces of F_2-folds (Fig. 5, point J). The most recent apparent motion along some of these faults is in a normal sense.

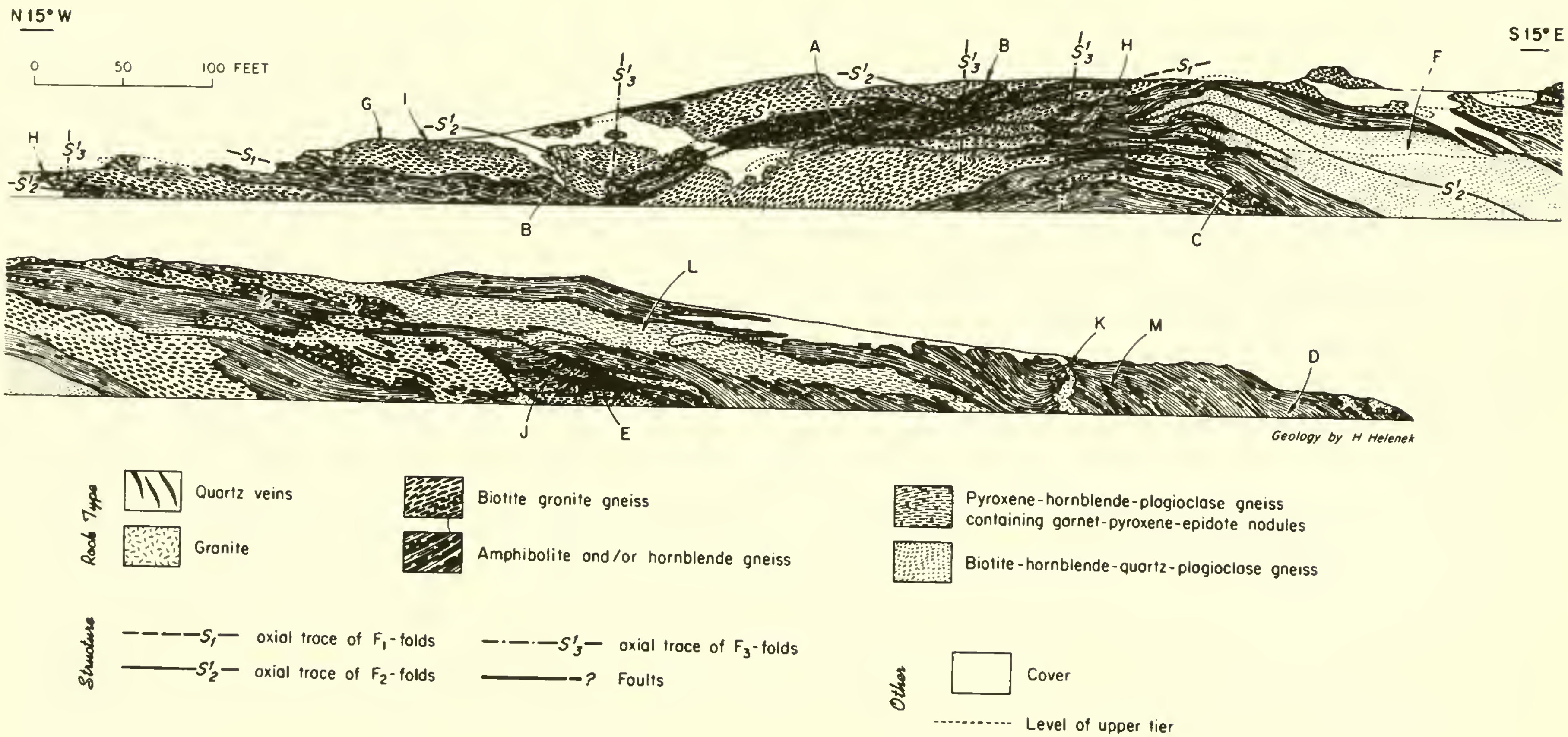
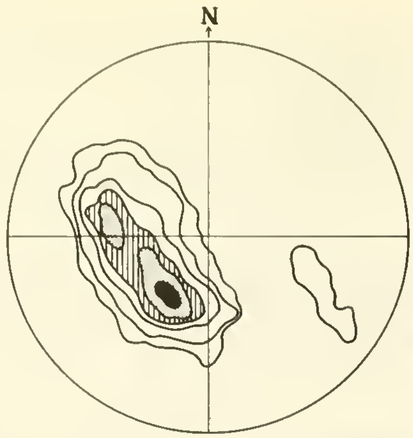


Figure 5. Rock-cut on the westbound lane of I-84 near Lake Carmel, Stop 2. Letter symbols are referred to in the text.



A. POLES TO COMPOSITIONAL LAYERING (S_0)

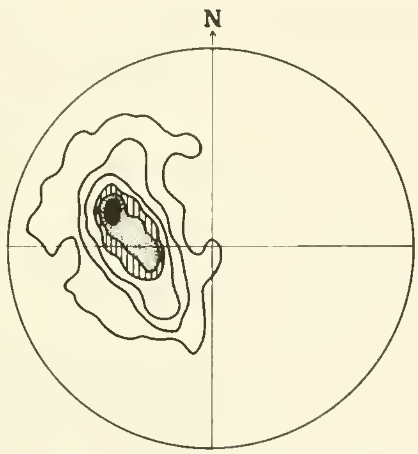
N = 622

1, 2, 4, 6, 8, 10% CONTOURS

B. MINERAL LINEATIONS (L_M)

N = 130

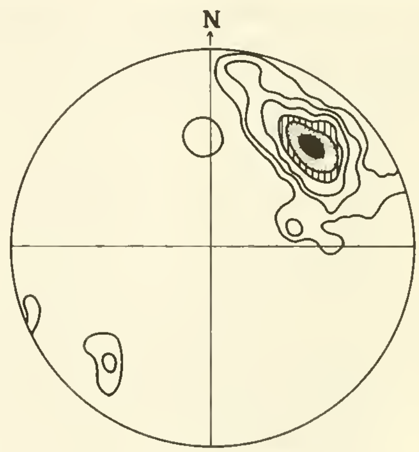
1, 3, 6, 9, 12, 15% CONTOURS



C. POLES TO F_2 AXIAL PLANES (S'_2)

N = 236

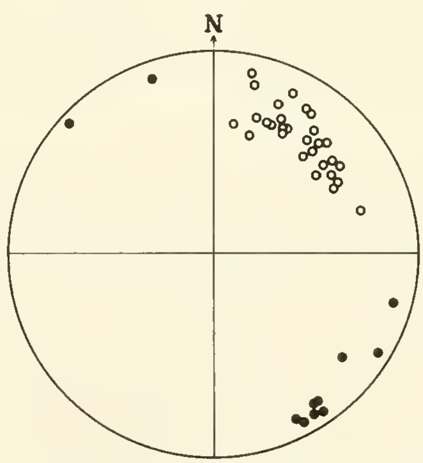
1, 3, 6, 9, 12, 15% CONTOURS



D. F_2 FOLD AXES (B'_2)

N = 71

2, 4, 8, 12, 16, 20% CONTOURS



E. ∘ 29, F_3 FOLD AXES (B'_3)

• 11, POLES TO F_3 AXIAL PLANES (S'_3)

Figure 6. Equal-area diagrams of structural data from the rock-cut on the westbound lane of I-84 near Lake Carmel, Stop 2.

Compositional layering (S_o) is transposed into an axial plane foliation (S_2) in the vicinity of the thrust faults (Fig. 5, point E).

A mineral lineation (L_M) (Fig. 6B) is prominent toward the north end of the exposure (Fig. 5, point G), but the significance of this lineation is not clearly understood.

Mineral assemblages in the garnet-pyroxene-epidote nodules (Fig. 5, point L) indicate that amphibolite facies metamorphism accompanied the F_2-folding. Radiometric data indicate metamorphism accompanying F_2-folding occurred during the Taconic orogeny.

Fabric elements of F_2-folds have been deformed into open upright F_3-folds. These F_3-folds are the gentle warps that refold the F_2-folds (Fig. 5, point H). Poles to the axial surfaces of F_2-folds lie on a girdle defining an axis oriented N55°E, 28°N (Fig. 6C) which coincides with the fold axes (B_3') of F_3-folds (Fig. 6E). The nearly coaxial relationship between the F_2- and F_3-folds suggests, but doesn't prove, that they were formed during one deformational event.

A final stage of folding, F_4, involves the warping of F_2 and F_3 fabric elements around a horizontal axis trending about N20°W. This warping has resulted in the reversal of plunge of early folds and is responsible for the dome-and-basin interference patterns on the nearly horizontal surface of the upper tier at the north end of the exposure (Fig. 5, point I). Post tectonic granite (Fig. 5, point K) and undeformed quartz-filled extension fractures (Fig. 5, point M) are present toward the south end of the exposure.

35.7		Take care when leaving the shoulder of the highway in returning to the westbound lane of I-84.

35.7		Alternate route to Stop 4: Proceed to the first exit ramp (approximately 1 mile) and leave I-84 in order to cross the interstate and return to it on the eastbound entrance ramp. Proceed east to Danbury, Connecticut and continue according to the main road log at mileage 88.3.

Continuation of Main Road Log

40.3	4.6	Cross the county line between Putnam and Dutchess counties and note the large exposure of Reservoir Granite. This is a sampling locality for Rb/Sr whole-rock dating (1296 ± 18 m.y.) of the Reservoir Granite.
43.2	2.9	This rest area on I-84 provides a very scenic view of the Hudson Valley. We will stop here if time permits.
45.6	2.4	This topographic break marks the boundary between the subdued topography of the Cambrian and Ordovician terrane and the more rugged Precambrian terrane.
45.8	0.2	Cross the Taconic State Parkway and continue west on I-84. There is a good view of the valley controlled by the Ramapo-Canopus fault zone to the left (south).
47.8	2.0	Bear right onto the exit ramp for exit 15, Lime Kiln Rd.
48.2	0.4	Bear right (north) onto Lime Kiln Rd.
49.3	1.1	Divided highway ends here. Continue approximately 0.1 mile and turn left (west) onto Route 52 and proceed westward.

51.0	0.8	Bear left (south) at the hamlet of Wiccopee onto Fishkill Hook Rd. and proceed approximately 0.1 mile, or less, and bear left and proceed southward on Fishkill Hook Rd.
52.1	0.4	The road forks here; bear right (south) onto West Hook Rd. and proceed.
52.8	0.7	Turn right (west) onto Van Wyck Lake Rd. Note the exposures of Poughquag Quartzite on the left.
53.4	0.1	Turn left into entrance to Sharpe Reservation and proceed straight ahead. In approximately 0.1 mile, drive through the gate.
53.7	0.3	Continue straight ahead past the road intersecting from the left.
53.8	0.1	Bear to the right where the side road intersects from the left.
53.9	0.1	Continue straight ahead, up the hill and pass intersection of side road on right.

54.0 0.1 Stop 3. Unconformity at Deer Lake (and lunch). This locality, where the Poughquag Quartzite rests unconformably on Precambrian gneisses typical of the Western Highlands, is on the grounds of Sharpe Reservation of the Fresh Air Fund.

Here the basal portion of the Poughquag Quartzite consists of an arkosic conglomerate, approximately 1 meter thick, overlain by quartzite containing lenses of subarkose and quartz pebble conglomerate. The arkosic conglomerate is drab, olive-gray to dull yellowish-white and consists of angular clasts of quartz and microcline in a sericitic matrix that contains minor amounts of zircon and highly altered biotite. The arkosic conglomerate at Deer Lake resembles the Precambrian Canada Hill type granite slightly. However it undoubtedly maps out as a stratigraphic horizon basal to the quartzite (Fig. 7). *Scolithus* is restricted to the quartzitic facies of the Poughquag.

Hornblende granitic gneiss, migmatitic biotite gneiss and amphibolite are the Precambrian rocks present here. The gneisses are finer grained and more highly altered than equivalent rocks in the Bear Mountain area probably due to the extensive brittle deformation that has occurred here. Quartz and feldspar are strained, hornblende is totally or partially replaced by green mica (biotite?), epidote, actinolite (?) and opaques. Biotite contains radiating acicular inclusions and is totally or partially replaced by chlorite, epidote, white mica and opaques.

The Poughquag Quartzite has been deformed into a series of open, symmetrical, folds, plunging gently northeast. Axial surfaces trend about N46°E and dip 80° NW and fold axes plunge N45°E, at less than 12°. Northeast of this locality the axial surfaces of these folds gradually change in trend toward the north and change in dip toward the southeast. Southeast of Wiccopee, about 1 mile north of Stop 3, folds in the Wappinger Group have axial surfaces oriented about N7°E, 58° SE and fold axes that trend N25°E, and plunge 14° NE.

Plastic deformation and at least one phase of brittle deformation have effected the Precambrian gneisses here but folds related to plastic deformation remain undetected in these exposures. Here and in immediately adjacent areas steeply dipping minor faults and microfractures occur in the gneisses. Most of the minor faults are strike-slip faults and the majority have a left-lateral sense. N52°W, N10°W and N27°E are the three prominent kinds of strike-slip movement. Quartz veins here are probably related to Paleozoic deformation.

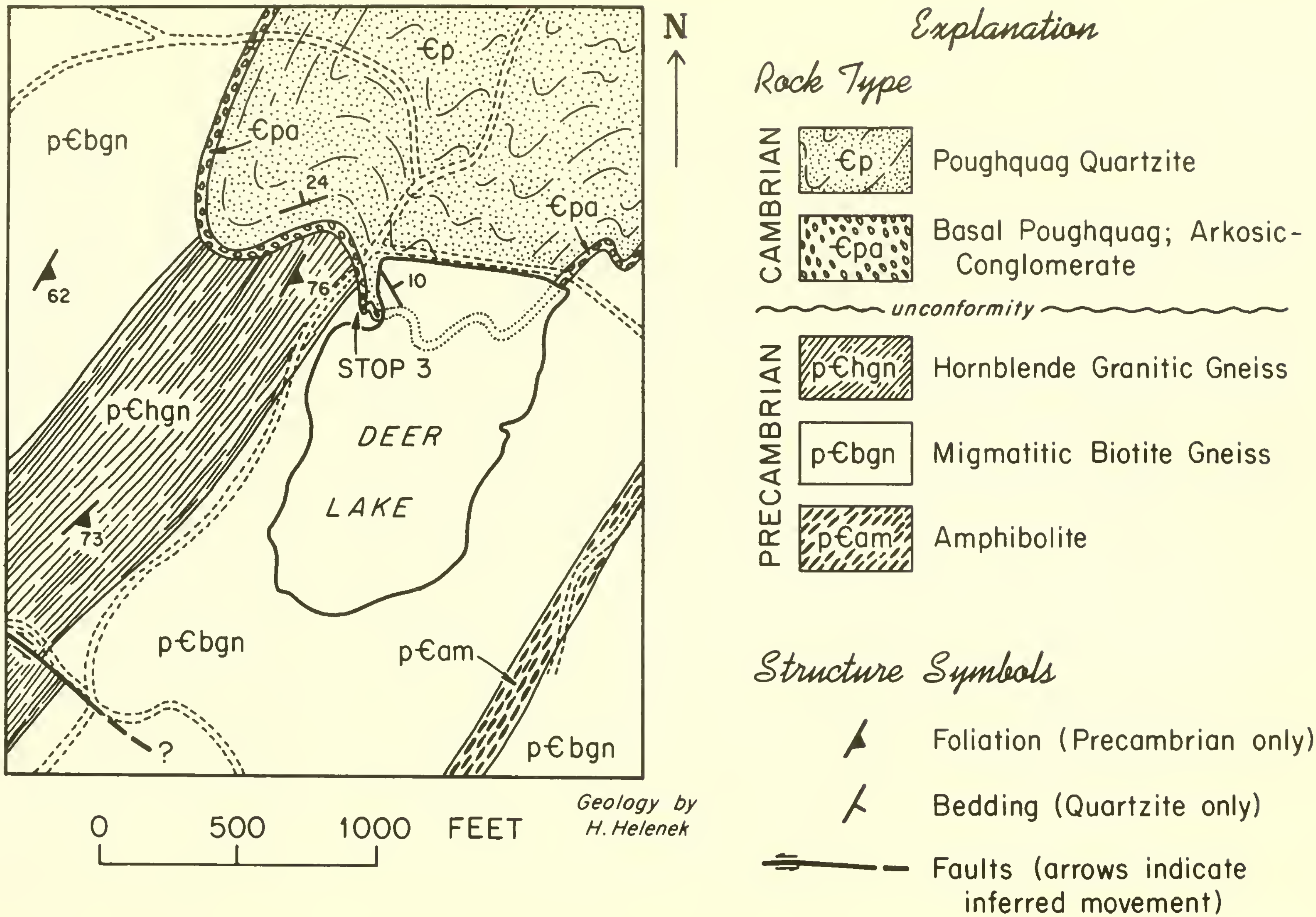


Figure 7. Geologic map of the region in the vicinity of Stop 3 on Sharpe Reservation.

54.0	-	Retrace route to the entrance to Sharpe Reservation.
54.6	0.6	Turn right onto Van Wyck Lake Rd. and proceed eastward toward West Hook Rd.
55.2	0.6	Turn left onto West Hook Rd. and in approximately 0.7 mile bear left onto Fishkill Hook Rd. at the junction of West Hook and East Hook Rds.
56.8	1.6	Bear right in the hamlet of Wiccopee and proceed a short distance to Route 52. Turn right (east) on Route 52 and retrace route back onto I-84.
58.6	1.8	Turn right (southeast) onto Shenandoah Rd.
58.7	0.1	Bear right (south) onto Lime Kiln Rd. (divided highway). Proceed south and cross overpass (I-84) in approximately 1.1 miles.
60.0	1.3	Enter the eastbound lane of I-84 and proceed to Danbury, Connecticut.

60.0	-	Alternate route to Stop 2: Take exit 19 at Route 312 and then proceed according to the main road log starting at mileage 32.8.

Continuation of Main Road Log

88.3	7.3	Stay toward the left and exit from I-84 onto Route 7 north.
89.1	0.8	Turn right onto Route 7 north and proceed to New Milford, Connecticut.
93.1	1.4	Woodville Marble is exposed in the road cuts being made in conjunction with the construction of new Route 7. Continue north on Route 7 toward New Milford.
100.3	7.2	Turn right (east) onto Route 67 and proceed across the Housatonic River. Keep to the left and go through the first traffic light and cross the railroad tracks. Turn left (north) onto Railroad St., at the second traffic light.
100.8	0.3	This is a complicated intersection with Bennett St. but continue essentially straight through it and onto Wellsville Ave. (Route 129).
102.4	0.3	Bear left at the fork onto Merryall Rd. (Route 129) and proceed northeasterly.
103.1	0.7	Turn left (west) onto Long Mountain Rd. and cross the West Aspetuck River.
103.6	0.5	Jog right and then left staying on Long Mountain Rd. at its junction with Aspetuck Rd.
103.8	0.2	Proceed straight ahead onto the dirt road (Bennett Rd.) at this junction.
104.2	0.4	There is a sharp curve to the left at the crest of this ridge and then a private drive on the left. Stay toward the right here and continue along Bennett Rd.

104.6 0.4 Stop 4. The "New Milford Massif" west of Long Mountain. A short traverse will be made from here along the road and then north to the vicinity of the power line.

Dark-gray to black-and-white, well-layered biotite-hornblende gneiss and thin beds of amphibolite constitute most of the Precambrian bedrock here. The Cambrian Lowerre Quartzite consists of brown-weathering sillimanite-microcline-biotite-quartz feldspar schist and schistose gneiss with prominent sillimanite-rich nodules,interbedded with feldspathic quartz granulites and slabby, gray- or tan-weathering quartzite. The base of the Lowerre is more siliceous and the sillimanitic schistose rocks predominate above. This is an example of the Lowerre Quartzite grading upward into rocks lithically similar to Manhattan C and is taken as further evidence that Manhattan C was deposited during the Cambrian and is an eastern facies of the Lowerre Quartzite (Hall, 1968 and 1971). Locality A on the traverse (Fig. 8A) is particularly striking because it clearly shows the basement gneisses physically on top of the Lowerre and the contact between the two units dipping gently northward. A large quartz-feldspar pegmatite is present near the power line and numerous inclusions of the country rock are present in the pegmatite near the contact.

The basement here occupies the core of an early isoclinal fold and the Lowerre is present on opposite limbs of the fold (Fig. 8A). The early isoclinal fold has been refolded as is clearly displayed by the sinistral map pattern of the Precambrian-Lowerre contact on both limbs of the early isoclinal fold (Fig. 8A). It is this later fold that accounts for the inverted Precambrian-Lowerre relationship that is present at locality A (Fig. 8A). Some of the rocks very close to the contact particularly at locality B display a cataclastic texture which is interpreted to be due to shearing along the unconformity during folding. The amount of transport associated with this shearing has not been determined but present indications are that it was not very large.

A prominent foliation and mineral lineation, that lies in the plane of foliation, developed during an early deformation. At the present time in the course of study of this area, it is not certain whether this foliation and lineation developed during Precambrian deformation or during the isoclinal folding involving the Lowerre Quartzite. The great circle spread of poles to foliation and the plunge of associated minor folds (Fig. 8B) indicate that the later sinistral fold plunges approximately N35°E at 20°.

The traverse at this stop will continue along the axial surface of the later antiformal fold toward the top of the hill where we will be able to see the Lowerre Quartzite physically on top of the basement. At this point we will be located on the eastern limb of the early isoclinal whereas locality A (Fig. 8A) is on the western limb of this early fold.

104.6	-	Return to vehicle, turn around, and return to Long Mountain Rd. by travelling south and east on Bennett Rd.
105.4	0.8	Junction of Bennett Rd. with Long Mountain Rd. and Vista Rd. Proceed straight ahead on Long Mountain Rd.
105.6	0.2	Jog right and then left, staying on Long Mountain Rd. at the junction with Aspetuck Rd.
106.1	0.5	Turn right (south) onto Merryall Rd.
106.8	0.7	Make an extremely sharp left turn (north) onto Paper Mill Road and proceed northward. The gneisses of the Bear Hill Massif are on the left (west) and the Lowerre Quartzite and Woodville Marble are toward the east.
108.8	0.2	Turn into the Eliot D. Pratt Education Center parking lot on right.

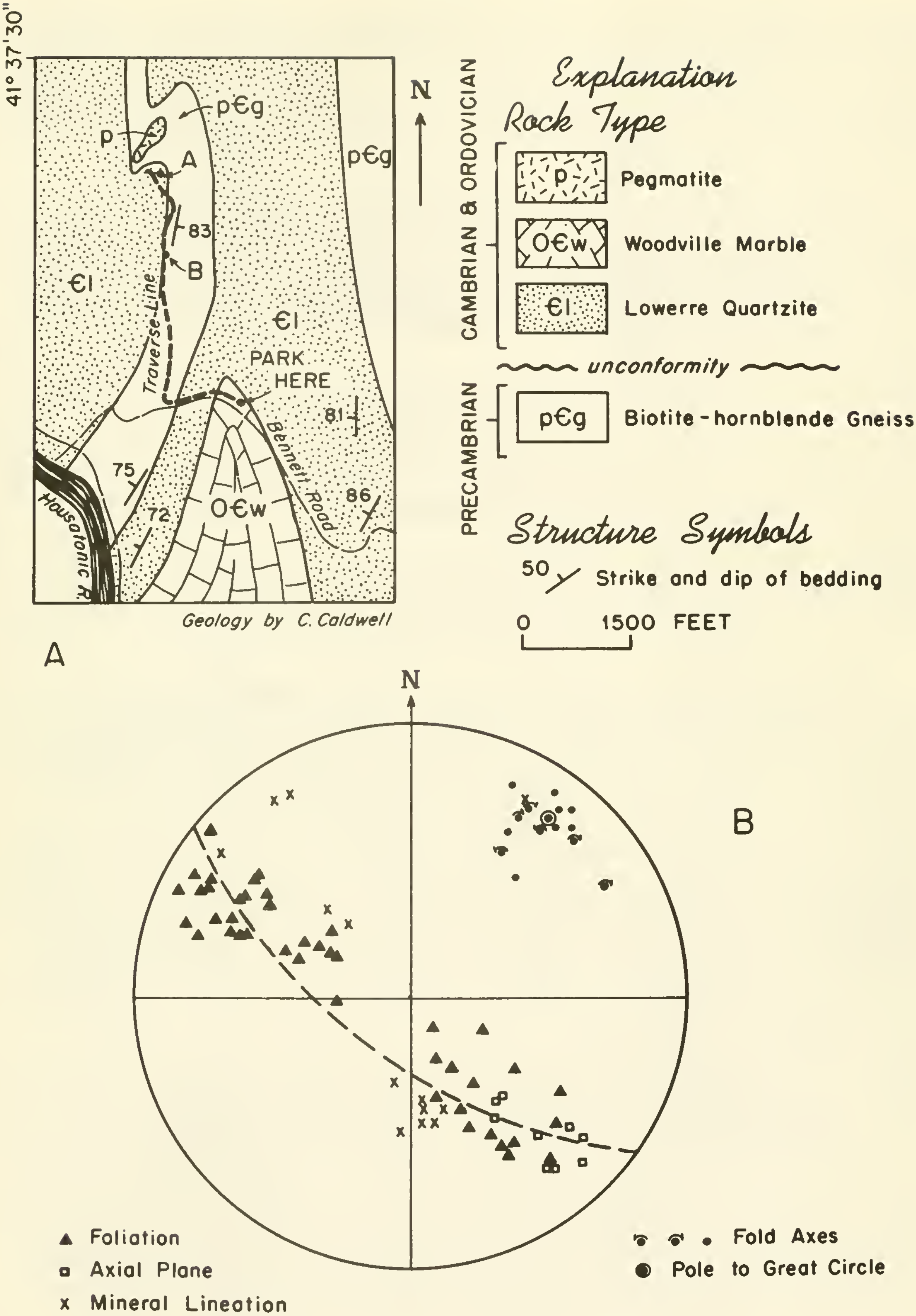


Figure 8. Geologic relations in the vicinity of Stop 4 are shown on the geologic map (A) and equal-area diagram (B).

Stop 5. Eastern boundary of the Bear Hill Massif. This ridge is underlain mainly by Precambrian rocks in the core of the Bear Hill Anticline with Cambrian Lowerre Quartzite on the east slope (Fig. 9). Four mappable Precambrian rock units have been defined in the area (Fig. 9) and Cambrian to Ordovician(?) carbonate rocks are exposed further east in the East Aspetuck River valley. A short traverse will be made here to study the Precambrian rocks and the Lowerre Quartzite.

Walk west along the road to the pasture entrance on the right. Ascend the east slope of the ridge past low outcrops of Lowerre Quartzite.
Station A (Fig. 9) - Several outcrops of well foliated light-pink to pink, biotite-quartz-feldspar gneiss with local pink microcline augen are present here. The abundance of biotite and degree of development of foliation are directly related and the foliated appearance is varied because the abundance of biotite differs from place to place.
Station B (Fig. 9) - Descend to the rock exposure on the east side of the terrace. This is the lower portion of the Lowerre Quartzite which lies unconformably on the Precambrian gneisses. It consists of well-foliated, reddish-weathering, feldspar-mica-quartz granulite that has thin laminae (1/8 to 1/2 inch thick) of quartzite and quartzose schist interbedded with well-foliated, gray- to tan-weathering, quartz-feldspar schistose granulite with nodules of quartz and feldspar (1/4 to 1/2 inch across) and minor beds of gray-weathering, massive micaceous and feldspathic-quartzite.
Station C - Walk northeast, along the unconformity approximately 800 feet past several rock exposures in the gray, biotite-hornblende gneiss and dark-gray, well foliated, amphibolite (p€h). Both contacts of this unit (p€h) are truncated by the unconformity at the base of the Lowerre (Fig. 9).

Structural data were collected at this station, approximately 75 feet west of the unconformity (Fig. 10A). Two phases of minor folds are present and the earlier of these deforms a foliation that is parallel to the compositional layering in the rocks. The foliation parallel to compositional layering very likely formed during a phase of folding that preceded the early folds that are so obvious here. The axes of the minor folds have an average plunge of 50° southwesterly and some of the folds have a counterclockwise rotation sense but the shear sense of most is indeterminate (Fig. 10A). The axial planes trend northeast and dip steeply northwest (Fig. 10A). These earlier folds are thought to be related to the large, map-scale fold, in the Precambrian units, which is truncated at the unconformity nearby (Fig. 9, point A). Crinkle folds that deform earlier features at this exposure and other nearby points represent a later phase of deformation. The wide variation in plunge of the crinkles is probably due to the varied attitude of previously folded foliation upon which the crinkles were formed.
Station D (Fig. 9) - Proceed east crossing a small stream and ascend the small ridge where pink quartz-feldspar gneiss (p€p) is present a few feet west of the unconformity and a deeply weathered zone is in the gneiss adjacent to the unconformity. Well-bedded Lowerre Quartzite, similar to that at Station B, is east of the unconformity here. Continuing up the ridge, the exposures of quartzite in the first 50-75 stratigraphic feet are predominantly thinly laminated granulite and quartzose schist with quartz-feldspar nodules. The stratigraphic section continues upward in the Lowerre to the east, over the ridge crest, and down the hillside. The thinly laminated beds and quartzose schists are less abundant with the dominant rock type being the massive gray- to tanish-gray-weathering, mica-feldspar quartzite, with local thin quartzite laminae, and feldspathic quartzite.

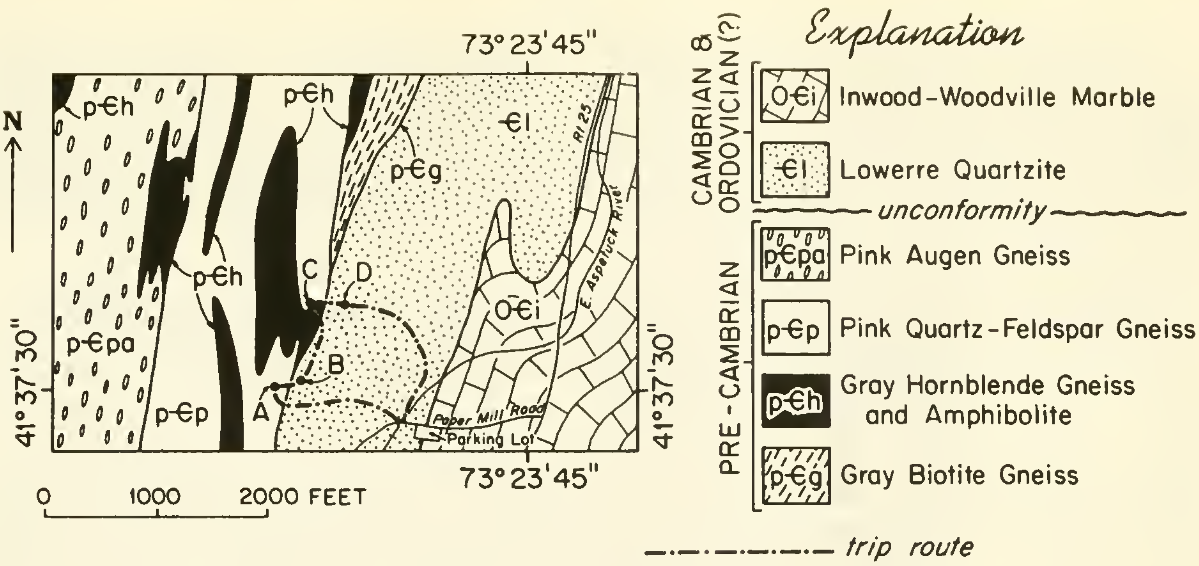


Figure 9. Geologic map of the region in the vicinity of Stop 5.

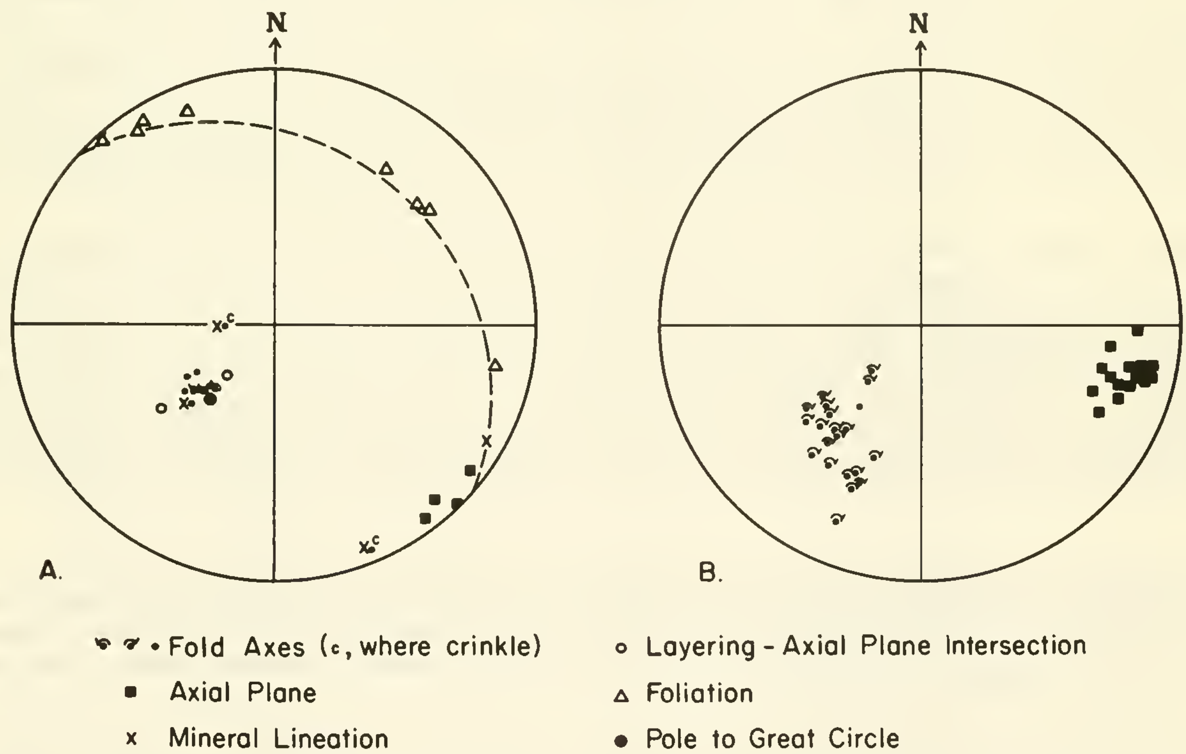


Figure 10. Equal-area diagrams that summarize structural data from the basement (A) and from the Lowerre Quartzite (B) at Stop 5.

One phase of minor folds is present in both foliation and bedding in the Lowerre (Fig. 10B). These folds have a clockwise rotation sense and an average plunge of 45° southwesterly. The axial planes trend north-northeast and dip steeply northwest (Fig. 10B). The minor folds of this phase of deformation, which are located on the east limb of the Bear Hill Anticline, have the proper rotation sense for an anticline to the west and thus are assumed to be genetically related to the large anticline.

Return to the starting point via a path near the East Aspetuck River.

108.8	-	Leave the parking lot and continue on Paper Mill Rd. to the junction with Route 25.
109.0	0.2	Turn left (north) onto Route 25 and proceed toward New Preston.
111.6	2.2	Note the road cut in the Lowerre Quartzite on the left (west).
113.1	1.5	Turn left (west) onto Route 45 toward Lake Waramaug and proceed through the village of New Preston on Route 45.
113.7	0.6	Stop sign is at the intersection at the south end of Lake Waramaug. Go through the intersection bearing to the right and continue on Route 45 to Warren.
116.9	0.6	Stop sign where Route 45 joins Route 341, bear left onto Routes 341 and 45 and proceed westward.
118.5	1.6	Bear left at the fork in the highway and proceed a short distance to the traffic light in Warren. Continue through this intersection proceeding westward on Route 341 to the village of Kent.
126.3	4.9	Bear to the right staying on Route 341 at this road junction and proceed toward the village of Kent.
126.9	0.6	Traffic light at the junction of Route 7 with Route 341 in Kent. Proceed straight through this intersection on Route 341 and cross the Housatonic River in approximately 0.2 mile.
127.4	0.5	Turn left into the parking lot for the Kent School ice-skating rink.

Stop 6. Kent School Road Cut. The road cut, located along the east edge of the Housatonic Highlands (Fig. 1),is in Precambrian gneisses that are overlain unconformably by the Cambrian Poughquag Quartzite. Two Precambrian rock units, pinkish-augen gneiss and interlayered gray, biotite gneisses, hornblende gneisses and amphibolite, are exposed in the cut. The augen gneiss is strikingly similar to pink augen gneisses near Stop 5 (Figure 9, p€pa) in the Bear Hill Massif.

The augen gneiss is a well-foliated pink and pale-pinkish-gray, biotite-quartz-plagioclase-microcline augen gneiss with widely scattered concentrations of garnet. The pale-pinkish-gray augen gneiss contains more plagioclase. Minor thin, dark-gray biotite gneisses with hornblende and epidote are locally interlayered with the augen gneiss.

The gray gneiss and amphibolite unit includes five main rock types, all of which are penetrated by granitic layers: well-foliated, gray hornblende-biotite-quartz-plagioclase gneiss, fine-grained, siliceous, gray, biotite-plagioclase-quartz gneiss, dark-gray, garnet-hornblende-quartz-biotite-plagioclase gneiss, dark-gray biotite amphibolite, and light-gray calc-silicate-rocks.

The Poughquag Quartzite consists of interbedded light-brown or buff-weathering quartzite and coarse-grained conglomeratic quartzite. Deeply weathered micaceous,conglomeratic-quartzite 3-5 feet thick is present at the base of the Poughquag and appears to be sheared. The Precambrian augen gneiss underlies this conglomerate and the higher massive quartzite beds of the Poughquag.

All of the rocks have been deformed by folding and faulting. Minor folds representing at least two stages of deformation are displayed in the Precambrian rocks and at least one stage of folding is represented in the Poughquag Quartzite. Faults are prominent particularly in the gneisses and a great deal of shearing is evident at the Precambrian-Poughquag contact.

Minor structural features measured at this road cut are shown on the three equal area plots in Figure 11. Structural data recorded from the Precambrian rocks on the north side of the road (Fig. 11B) show numerous fold axes clustered about the pole to the great circle defined by poles to foliation. These are the axes of the earlier of two sets of folds present in the Precambrian here and they plunge S22E at 42° (Fig. 11B). Trends of these fold axes are scattered from S07E to S40E, and the associated axial plane foliation strikes from N20W to N20E and dips 70°-80° easterly. Foliation that is parallel or subparallel to the compositional layering,also folded by this deformation, probably formed during an even earlier deformation. Poles to this foliation consititue a well-defined great circle and beta maximum. Crinkles deform both the compositional layering foliation and the axial plane foliation of the southeast plunging folds. These crinkle axes trend from S26W to S58W and plunge gently to moderately southwest (Fig. 11B). Biotite lineation trends from S35E to N85E and plunges moderately eastward. The biotite lineation appears to be deformed by the southeast plunging folds and therefore to have formed during the same event that produced the prominent foliation that is parallel to layering. Several fairly good candidates for early isoclinal folds associated with the lineation and foliation are on the south side of the road. Several faults strike northeasterly and dip moderately southeast.

Earlier phase folds which deform the compositional layering foliation and mineral lineation have a somewhat different orientation on the south side of the road (Fig. 11C). With the exception of one fold axis trending ESE, these fold axes on the south side of the road trend from S02E to S08E and plunge moderately south. Poles to the layering foliation lie on a well defined great circle, the pole to which is parallel to the axes of earlier phase minor folds (Fig. 11C). Many faults and shear zones with associated granitic rocks are present on the south side of the road and most trend northeast to east-northeast and dip moderately to steeply southeast (Fig. 11C).

Bedding in the Poughquag Quartzite south of the road dips moderately southeast and strikes from N30E to N80E (Fig. 11A). A poorly defined cleavage is locally present and is subparallel to bedding and one poorly defined minor fold has been found in the Poughquag at this locality. It was not possible to accurately measure the fold axis directly but it clearly plunges moderately southeastward. The axial plane of this fold strikes approximately N37E and dips 39SE and the intersection of the axial plane with bedding is S30E, 37SE (Fig. 11A). Quartz and tourmaline lineations near this fold plunge S57E at 31° and S47E at 20°. Three prominent joint sets are present in the quartzite (Fig. 11A). The most prominent set trends northeast and dips moderately northwest, another set trends N10W and is nearly vertical while the third set trends approximately N45W and is

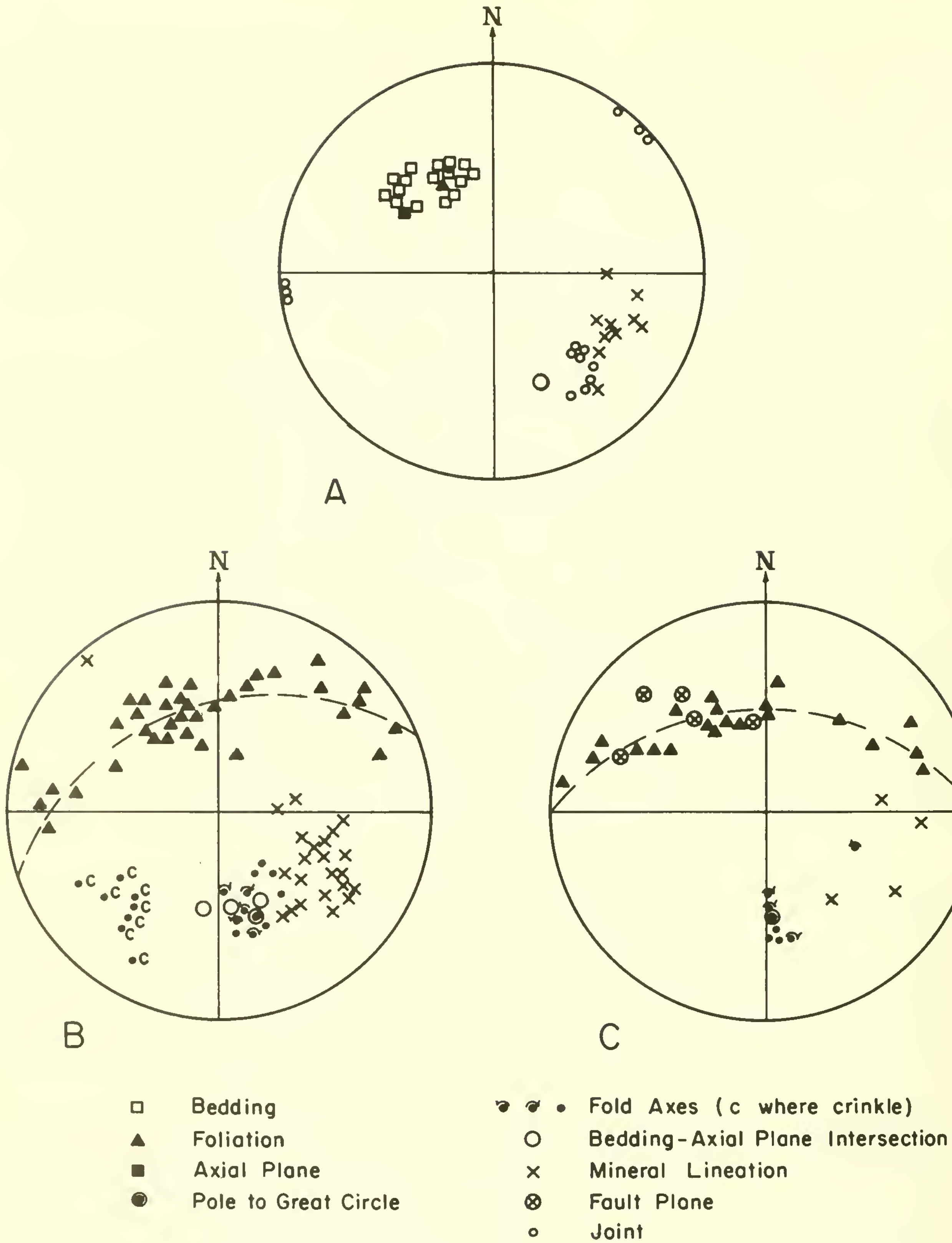


Figure 11. Equal-area diagrams summarizing structural data from the Poughquag Quartzite (A) and from the basement north of the road (B) and south of the road (C) at Stop 6.

nearly vertical.

A three to five foot thick zone of shearing is evident near the Precambrian-Poughquag contact. Both the augen gneiss and micaceous quartz conglomerate were involved in this shearing as indicated by the cataclastic texture and deep weathering of both rocks in this zone. The extent of movement along this zone is not certain at the present time in the on going study of the Kent quadrangle. It may represent minor local shearing along the contact between rocks of contrasting ductility during folding or it may represent a major, more regional, fault where the Paleozoic cover has sheared over the basement.

127.4	-	Leave the parking lot and turn right (east) onto Route 341. Proceed across the Housatonic River to the traffic light at the junction of Route 7 with Route 341.
128.0	0.6	Turn left (north) onto Route 7 and proceed north on Route 7 to Great Barrington, Massachusetts. The trip from Kent to Great Barrington takes approximately one hour,or a little less, depending on traffic conditions.

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TRIP A - 2: THE HUDSON ESTUARY

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The Hudson Estuary Survey

The City Institute of Marine and Atmospheric Sciences' Hudson Survey program began in October, 1973 with a series of initial cruises to establish a viable survey track and sampling protocol. A series of 23 stations (Fig. 1, Table 1) were selected as the survey track and these stations have been sampled once per month. The entire survey area covers a linear distance of approximately 115 mi extending from Saugerties, New York on the north (station 1) to a southernmost station (16B), 3.5 mi northeast of Sandy Hook, New Jersey, at the apex of the New York Bight. Since the survey represents a multidisciplinary effort, the following sampling activities were performed at each station:

A. Precision depth recording
B. Three bottle hydrocast to take samples from the surface, mid-depth, and bottom of the water column.
C. Bottom sediment collection using a Peterson grab sampler.
D. Phleger core sampling.
E. Secchi disc transparency.
F. Salinity, temperature, and dissolved oxygen determinations at every two meters of depth from surface to bottom.

The effort to date represents over 58 actual ship days on the estuary during 16 cruises on the Institute's 88 ft research vessel the R.V. Commonwealth. The following is a series of papers presenting some of the preliminary results of this survey in the broad areas of the physical, chemical, geological, and paleontological observations obtained during the first eighteen months of study.

Geologic Setting and History of the Hudson Estuary

The Hudson River and Estuary is one of the major waterways of the eastern part of the United States. It begins in the southern part of the Adirondack Mountains and flows southward approximately 300 km to its mouth at New York City. About 34,650 km^2 of the southeastern part of New York, northeastern New Jersey, and southwestern New England comprise the watershed of the Hudson River and Estuary.

The depth of the estuary in the area of this trip remains relatively constant. From south of The Highlands to The Narrows and from north of The Highlands to Poughkeepsie, water depths average 15 to 16 m. It is in the gorge of the Hudson Highlands that the river abruptly reaches depths of as much as 55 m. Both in The Highlands and north to Poughkeepsie, the main channel of the estuary covers the full width of the valley, which is between 1200 and 1500 m wide. South of The Highlands, in Haverstraw Bay and the Tappan Zee, broad expanses of shallow water border the main channel of the estuary. Here,

Contribution No. 46 of the City Institute of Marine and Atmospheric Sciences.

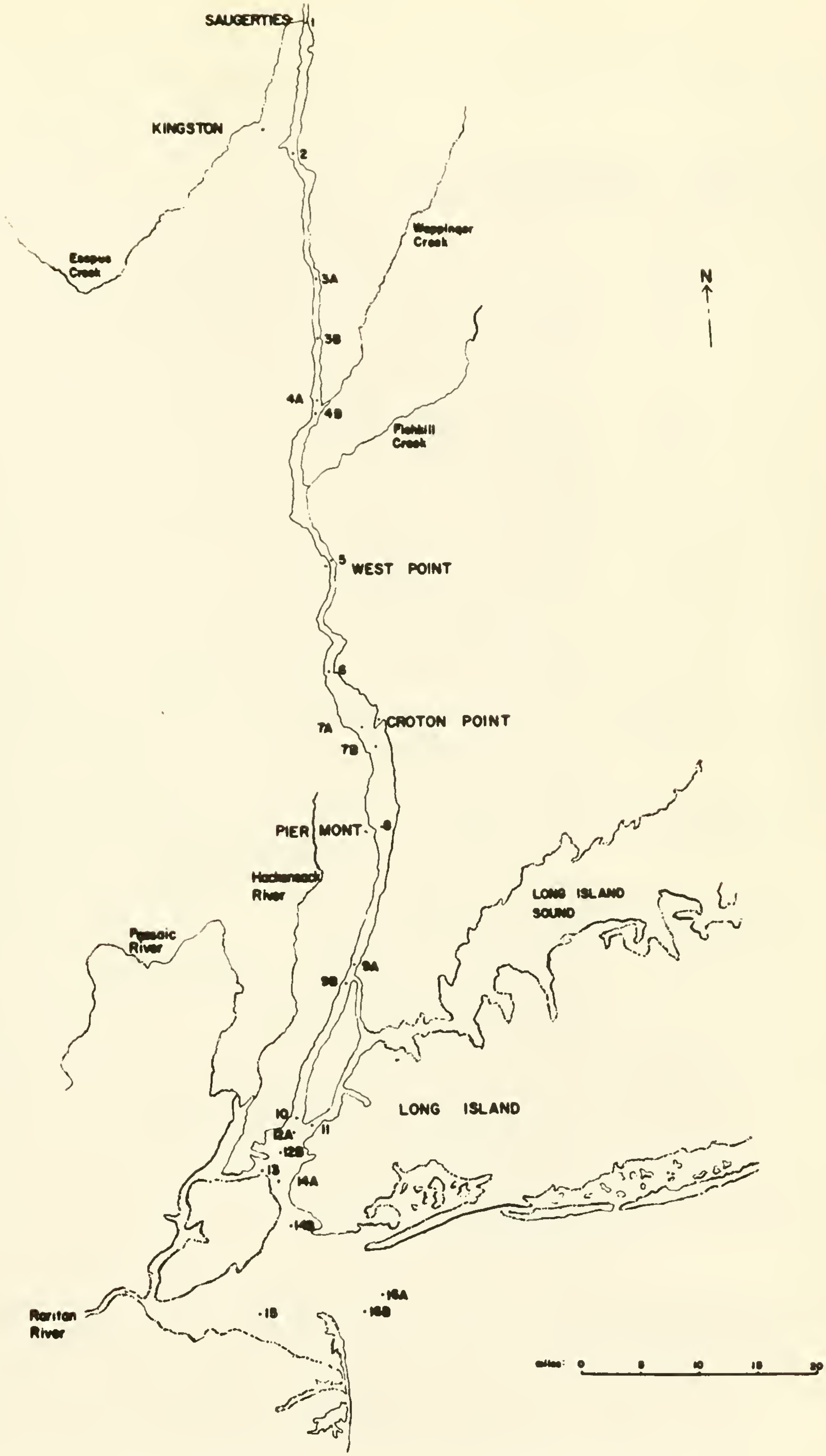

FIGURE 1. MAP OF SAMPLING SITES OF THE HUDSON ESTUARY SURVEY.

Table 1. SECTOR AND STATION LOCATIONS

Sector	Station	Location	Mile Point*
16	16A 16B	N.Y. Bight, 6.0 mi ne of Sandy Hook N.Y. Bight, 3.5 mi ne of Sandy Hook	-15
15	15	Lower Bay, 3 mi nw of Sandy Hook	-12
14	14A 14B	North of The Narrows South of The Narrows	- 8
13	13	Kill Van Kull	- 6
12	12A 12B	North end of Upper Bay South end of Upper Bay	- 2
11	11	East River, near The Battery	0
10	10	Hudson River, near The Battery	0
9	9A 9B	North of Spuyten Duyvil South of Spuyten Duyvil	+ 8
8	8	Piermont, New York	+18
7	7A 7B	North of Croton Point, New York South of Croton Point, New York	+25
6	6	Verplanck, New York	+33
5	5	World's End (West Point, New York)	+50
4	4A 4B	North of Wappinger Creek South of Wappinger Creek	+63
3	3A 3B	North of Poughkeepsie, New York South of Poughkeepsie, New York	+71
2	2	Kingston, New York	+88
1	1	Saugerties, New York	+99

* + north of The Battery; - south of The Battery

just north of Croton Point, the estuary reaches its maximum width of 5.5 km. Between Piermont and The Battery at New York City, the estuary is 1800 m wide and occupies the full width of the valley.

Upper New York Bay marks the confluence of the Arthur Kill and the East River with the Hudson. The constriction of the estuary's mouth (The Narrows) is caused by the morainal deposits (Harbor Hill) which cross the estuary at its southern end. The southernmost area, Lower New York Bay, marks the apex of the innermost part of the New York Bight.

As seen on this trip (Piermont to Poughkeepsie), the Hudson Estuary lies in a region of Precambrian to late Pleistocene rocks and sediments. A variety of igneous, sedimentary, and metamorphic rocks comprise the valley of the Hudson Estuary (Fig. 2).

The oldest rocks found in the Hudson Valley are in The Highlands. Granites and various kinds of gneiss are the dominant rocks of The Highlands. Radiometric dating indicates a Precambrian age of 1.2 b.y. (Grenville).

In the area south of the Hudson Highlands, the estuary flows along the contact between Triassic rocks comprising the Newark Group with the intruded Palisades Sill on the west and Precambrian to lower Paleozoic crystalline rocks on the east. The Newark Group consists of continental sedimentary rocks (primarily red-beds) that have been intruded by the diabase of the Palisades Sill which crops out as the conspicuous cliffs along the west side of the valley south of Stony Point, New York. The Precambrian and lower Paleozoic rocks cropping out on the east side of the valley south of Verplanck, New York consists of gneiss, schist, and marble of the New York City Group. In addition, the area between Stony Point and The Highlands on the west bank of the valley is underlain by the Cambro-Ordovician Wappinger Limestone. The Cortlandt Complex of Devonian age and lower Paleozoic inliers are found between Verplanck and The Highlands on the east side of the valley.

North of The Highlands, the Hudson River flows across folded and faulted Cambrian and Ordovician sedimentary rocks. These include the Poughquag Quartzite (Cambrian), Wappinger Limestone (Cambro-Ordovician), and the Normanskill Shale (Ordovician).

In New York Bay and The Narrows the estuary is developed on Cretaceous coastal plain deposits, and Pleistocene glacial sediments. Late glacial varved clays and fluvial sands and gravels, including the (former) deltaic deposits of Croton Point, underlie most of the Postglacial and Recent estuarine sediments that have been deposited in the Hudson.

The Hudson River is usually considered to have begun its development during the Cretaceous Period following the Fall Zone cycle of erosion (Johnson, 1931). According to Johnson, the river attained its present course and basic configuration in the Tertiary Period as a result of a series of stream captures. The present valley was modified during the various Pliestocene glaciations and also by Pleistocene lacustrine, fluvial, and estuarine processes. The late Pleistocene ice sheet not only deepened the bedrock channel of the river, but its subsequent retreat 17,000 to 18,000 years ago (Connally and Sirkin, 1970) contributed to a rise of sea level that drowned a major portion of the river and created the present estuary.

Following the draining of glacial Lake Hudson, tidal conditions were established in the estuary well before 12,000 yr ago (Newman _et al_., 1969). Estuarine conditions with salinities high enough to support foraminifers (approx. 30°/oo) became established by about 11,500 yr ago (Weiss, 1974). About 10,000 yr B.P., salinity decreased slightly but was re-established by 9,000 yr ago. The maximum transgression of mesohaline brackish water

A-2

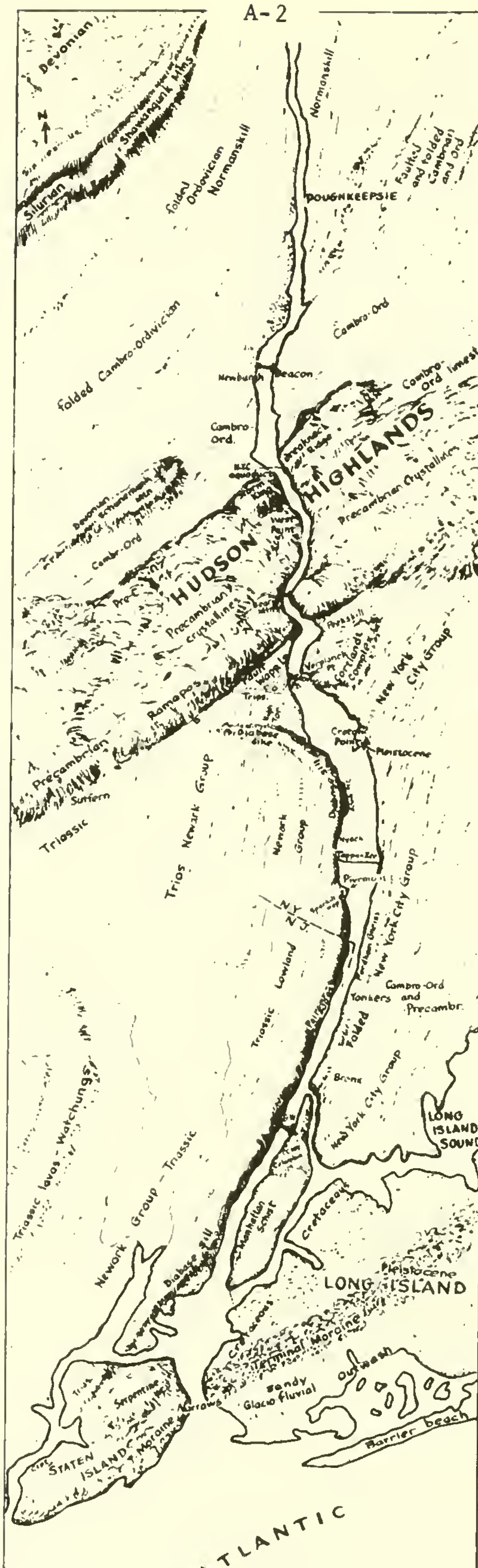

FIGURE 2. PHYSIOGRAPHIC DIAGRAM OF THE LOWER HUDSON RIVER VALLEY

(salinity between 5.0 and 32^{o}/oo) into the estuary during late- and postglacial time occurred about 6,500 yr B.P., as shown by the initial appearance of foraminifers in the Peekskill area of the estuary. This event coincides with the general flooding of the northeastern United States by the Atlantic Ocean. Foraminiferal evidence indicates that the salinity of the estuary has decreased during the past 1,500 to 3,000 years. This appears to be the result of sediment being deposited faster than the rise of sea level or crustal subsidence. Thus, the influence of saltwater within the estuary has been regressing.

Field Trip Itinerary

The Hudson Estuary field trips (A-2, Friday October 10, 1975; and C-12, Sunday October 12, 1975) will study six sectors of the Hudson, from sector 8, Piermont, New York to sector 3, Poughkeepsie, New York (Fig. 1, Table 1). The vessel will occupy six sampling stations on each trip; sectors 3, 4, 5, 6, 7, and 8. These stations will include sections of the estuary which display the shallow and wide bay areas (sectors 6, 7, 8), the deepest section (sector 5), a major tributary outfall (sector 4), and an area of dense urbanization (sector 3). The following onboard operations will be performed at each station:

1. Precision Depth Recording for depth and topographic profiling.
2. Salinity - Temperature - Dissolved Oxygen - Depth Profile.
3. Phleger core sampling.
4. Peterson grab sampling for bulk sediment and benthic biota.
5. Zooplankton tow primarily for icthyoplankton.

Water samples will be collected if requested by the participants. The significance of these sampling procedures and the results obtained at the stations to be occupied by this field trip are discussed in the accompanying papers in this section of the guidebook.

Trip A - 2 will steam on the R.V. Commonwealth from Piermont, New York at 8:00 A.M. and will arrive at Poughkeepsie, New York at approximately 6:00 P.M. Transportation will be provided from Poughkeepsie to the conference at Great Barrington, Vermont.

Trip C - 12 will depart Great Barrington at 7:30 A.M. and will steam on the R.V. Commonwealth from Poughkeepsie, New York at 9:00 A.M. Sampling operations will be completed at sector 8. From here the ship will proceed to St. George, Staten Island, New York arriving at approximately 10:00 P.M. Public transportation is available from Staten Island.

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TEMPERATURE AND SALINITY OBSERVATIONS IN THE HUDSON ESTUARY, 1974

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Introduction

As part of the City Institute of Marine and Atmospheric Sciences' ongoing multidisciplinary study of the Hudson Estuary, seasonal changes in the temperature-salinity regime of the study area were recorded. This paper reports data from January to November, 1974 measured over a survey track of approximately 115 mi in length, extending from Saugerties, New York on the north, to a southern position, in the apex of the New York Bight, at a point 3.5 mi northeast of Sandy Hook, New Jersey. This survey track included the estuarine section of the Hudson, as well as, freshwater and marine reference stations.

Methods Of Study

A series of 16 sampling sectors (Table 1), containing a total of 23 stations (Fig. 1) was established. Certain stations: 14A and 14B, at each end of The Narrows; 12A and 12B, at each end of Upper New York Bay; 9A and 9B, on either side of the confluence of the Hudson and Harlem Rivers; 7A and 7B, north and south of the mouth of the Croton River; 4A and 4B, north and south of the confluence of the Hudson River and Wappinger Creek; 3A and 3B, north and south of Poughkeepsie, New York, were taken to be tidal stations and were located such that the "A" stations were sampled on the flood tide, while the "B" stations were sampled on the ebb tide. All stations were in or adjacent to the main channel of the river and each sampling sector was occupied at least once per month during the sampling period.

Actual measurements of the Salinity-Temperature-Depth profile at each station was made using a deck lowerable _in situ_ conductivity, salinity, temperature, and depth sensor run concurrently with standard water sampling using Niskin bottles and reversing thermometers. These latter samples were taken as part of a continuing water sampling program including checks of salinity using an on board salinometer.

In order to produce sigma-t (σ_t) curves and gain an understanding of density stratification within the study area, the computer program of Cox _et al._ (1970) was used along with the anticipated and measured temperatures and salinities of the area. While this method does not take into account turbidity related density, it does represent a reasonable first approximation of Temperature-Salinity-Density relationships anticipated in the study area. If the actual Temperature-Salinity profiles of the Estuary are parallel to the generated sigma-t lines of equal density, then there is good presumptive evidence for a condition of at least partial salinity stratification.

This work was jointly sponsored by Lehman College and the City Institute of Marine and Atmospheric Sciences.

Contribution No. 47 of the City Institute of Marine and Atmospheric Sciences.

Results and Discussion

Figure 3 is a graphic representation of equal density profiles generated using the program of Cox et al. (1970), over the salinity-temperature regime of the study area. Figure 4 represents the Temperature-Salinity profiles of representative sectors during the winter (January), spring (April), summer (July), fall (October), and again winter (November) flow periods of the estuary. Comparison of the slopes of these Temperature-Salinity profiles with the sigma-t lines of equal density (Fig. 3) indicate that during all flow regimes of 1974 the estuary was partially stratified with more dense saline water riding below less dense saline water. This is in agreement with the findings of Abood (1974), who described the Hudson Estuary as a partially stratified estuary with a net seaward flow. Further, it will be noticed that the northward progression of high salinity water as a result of tidal forcing during low flow periods is significant and that detectable salinity concentrations were observed north of Haverstraw Bay and even above Verplanck (sector 6), some 35 mi north of The Battery (Fig. 1) in November, 1974.

The observations that the Temperature-Salinity (T/S) profile of station 8 (Piermont) in April is represented by a single point is indicative of total mixing of the water column at that time. Further, the slopes of the T/S profiles of stations 1 to 7 in April and October, and stations 1 to 5 in July and November approach the slope of the sigma-t lines for density (Fig. 3), indicating at least partial mixing of the water column at these stations during these months. The same reasoning and conclusions can be drawn for station 8 in October and stations 6, 7, and 8 during November, 1974.

It will be also observed from figure 4, that the slope of the T/S profile for station 15 during the months of January, October, and July is markedly different from the slopes of stations 9, 12, and 16 during the same months. This can be explained by the fact that during these months station 15 was sampled during an ebbing tide and at this time measurements were for water originating from Raritan Bay, draining from the Raritan River, the Arthur Kill, and Newark Bay, and clearly not water from the Hudson Estuary system. Although not shown in the data, the same observation was made for the waters of stations 11 and 13, which during the ebbing tide are derived from Long Island Sound and Newark Bay (via the Kill Van Kull) respectively, and are clearly not part of the linear tract of the Hudson Estuary system. However, on the flooding tide, as seen in figure 4, for station 15 during April and November sampling, and observed at stations 11 and 13 during the same periods, the T/S profiles "line-up" with the linear T/S representation for the Hudson Estuary. The observations have led to the design of future sampling tracks and time relative to ebb and flow of the tides which will help define the origins and mixing patterns of water types within the Hudson Estuary system and the apex of the New York Bight.

Finally, it will be observed (Fig. 4), from a temperature point of view, that the saltwater end of the survey track had a 13°C range during the sampling period while the freshwater end had a thermal range of 25°C. At any given month during the sampling period the T/S profile of the entire estuary can be approximated by a straight line connecting the saltwater input with the freshwater source, north of station 6, and representing a gradual mixing of these water types. This straight-line property of the Hudson Estuary water mass implies that during the sampling period surface heating, cooling, evaporation, and precipitation did not have significant effects on the properties of the Hudson Estuary water mass.

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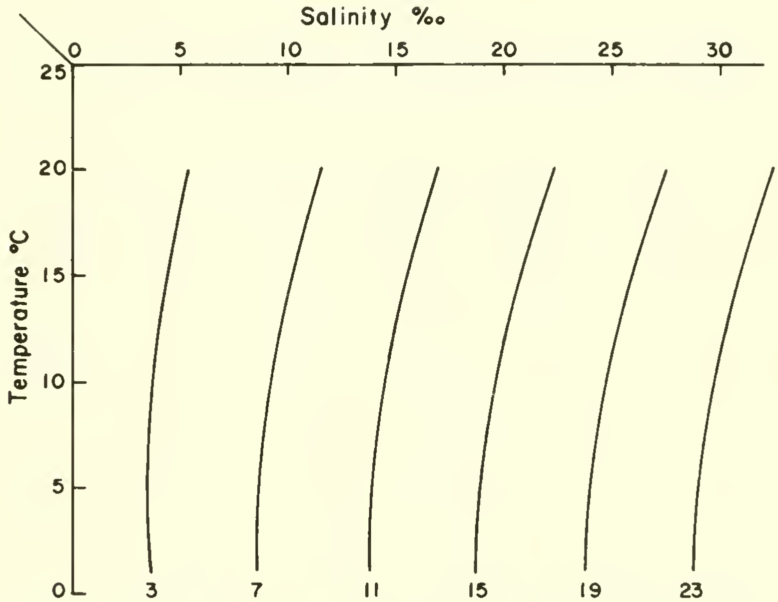

Fig. 3. σ_T after R.A. Cox, et al. 1970 Deep Sea Research. 17:679-689

$$\sigma_T = \left[\frac{\text{Density of Sea Water at } t°C}{\text{Density of Sea Water at } 4°C} - 1\right] 1000$$

$$\sigma = \sum_i \sum_j a_{i,j} T^i S^j \quad 0 \le i \text{ and } j \le 3 =$$

$$a_{0,0} + a_{1,0}T + a_{0,1}S + a_{2,0}T^2 + a_{1,1}ST + a_{0,2}S^2 + a_{3,0}T^3 + a_{2,1}ST^2 + a_{1,2}S^2T + a_{0,3}S^3$$

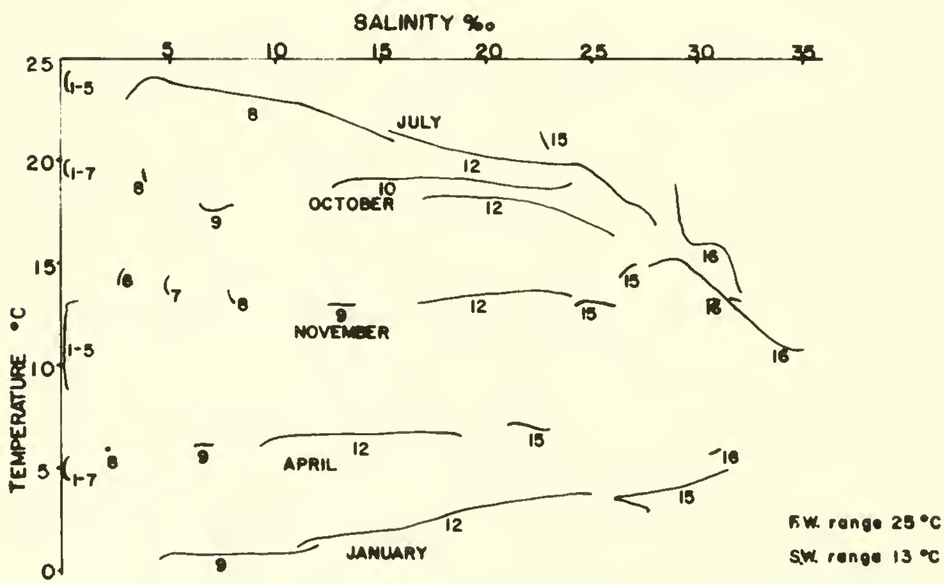

Fig 4. HUDSON ESTUARY T/S PROFILE

Jan. - Nov. 1974

OBSERVATIONS OF THE TRACE METAL LOAD OF THE HUDSON ESTUARY, 1974

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Introduction

As part of the City Institute of Marine and Atmospheric Sciences' Hudson Estuary study, soluble trace metal determinations were made from surface and bottom waters at each of the stations listed in Figure 1 and Table 1. The object was to record the changes, on a month to month basis, of the soluble trace metal load of the system and to observe the movements of these metals with respect to the tidal movement of the estuary.

Methods of Study

Surface and bottom water was collected in 8-liter Niskin (PVC) bottles at each station and a 500 ml sample from each bottle was filtered through a 0.45 micron Millipore filter to remove particulates. The filtered water was stored in acid washed polypropylene bottles and acidified to pH 2 using trace metal free concentrated HCl. These acidified filtered water samples were sent to the laboratory where they were analyzed for their lead, copper, and cadmium content. The analysis procedure involved triplicate runs using the technique of standard additions and was accomplished by the polarographic method, using an Environmental Sciences Associates, Inc. Anodic Stripping Voltammeter. This technique employes a mercury-graphite electrode for uniform geometry and requires that the samples be buffered to pH 5 using a sodium acetate buffer. The theory of this procedure and its application to aquatic environments has been documented by Allen, Matson, and Mancy (1970) and has the advantage of high sensitivity in the nanogram range.

Results and Discussion

Figure 5 presents the soluble trace metal data (Cu, Pb, Cd) for the months of March, May, August, and November, 1974. These months are representative of the spring, summer, and winter flow regimes of the estuary. It will be observed from the figure that the trace metal range over the whole water column and extent of the sampling track was as follows:

Month	Cadmium	Lead	Copper
March	0 - 3.45 ppb	0.2 - 6.9 ppb	2.0 - 26.0 ppb
May	1.0 - 7.6 ppb	1.0 - 4.0 ppb	3.2 - 13.2 ppb
August	0 - 0.8 ppb	1.3 - 3.4 ppb	5.3 - 15.0 ppb
November	0.3 - 1.1 ppb	1.1 - 4.0 ppb	5.1 - 13.4 ppb

Further, one can see that the metals occur in the water column such that the concentration sequence is Cu > Pb > Cd.

This work was jointly supported by Lehman College and the City Institute of Marine and Atmospheric Sciences.

Contribution No. 48 of the City Institute of Marine and Atmospheric Sciences.

By understanding the state of the tide at the time of sampling at each station one can speculate that metal inputs into the system occur at station 13, the Kill Van Kull (which is observed as metal peaks at station 12); between stations 9A and 9B (Harlem River); between stations 7A and 7B (Croton River); between stations 3A and 3B (Poughkeepsie); and station 1 (Saugerties). This is particularly evident by the peak Cu values at stations 12, 9, 7, 3, and 1 for all months, and peak Pb values at station 12 in March and November, as well as, the peak Cu, Pb, and Cd values at station 3 during May. There also appears to have been a Pb input into the system at Saugerties during the months of March and May.

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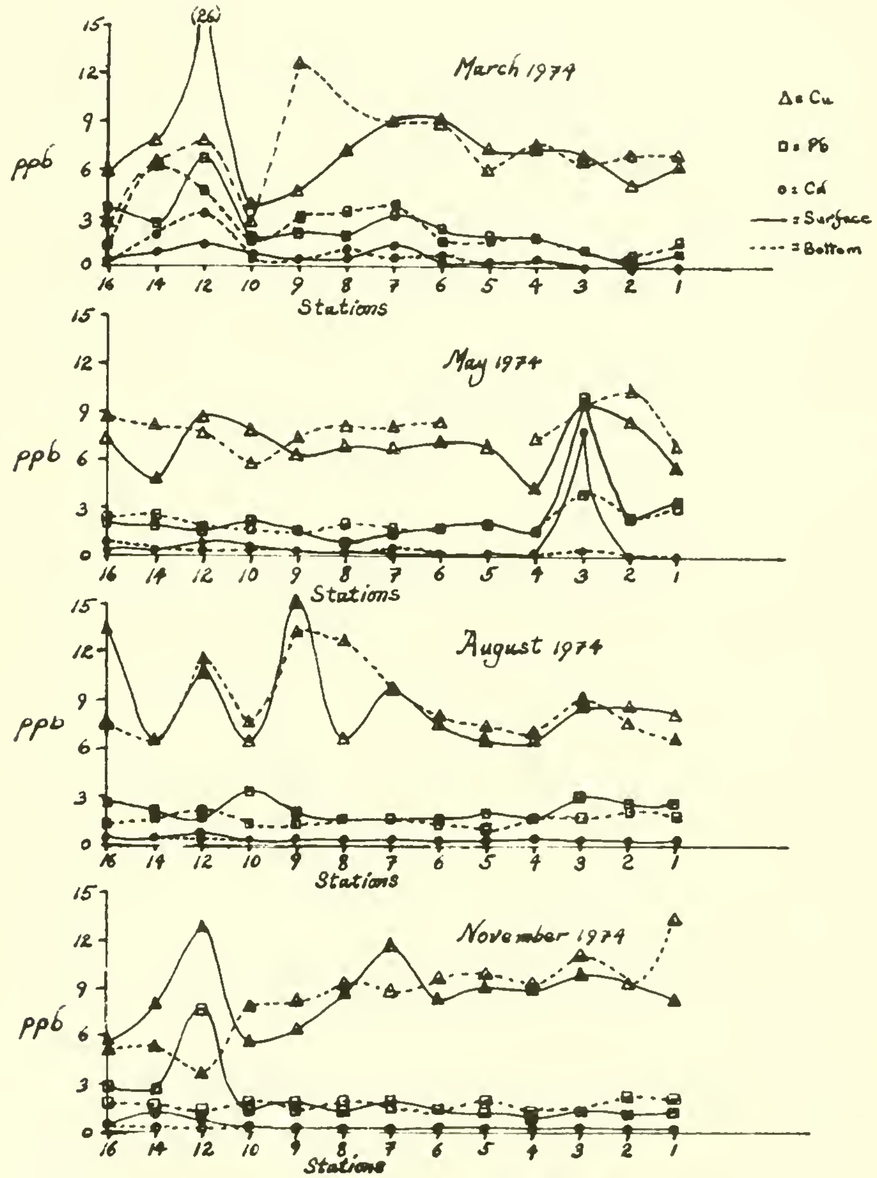

Fig. 5. Hudson Estuary Survey. Soluble Trace Metals

THE DISSOLVED OXYGEN IN THE HUDSON RIVER ESTUARY

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The dissolved oxygen present in a body of water is one indicator of the potential of that water to serve as a healthy biological environment. Many factors, such as the oxidative breakdown of organic materials present, respiratory activities of heterotrophic organisms and surface films of oil or detergents will reduce the total dissolved oxygen, which has entered the water by diffusion, or been produced there by photosynthetic organisms. The distribution of the dissolved oxygen throughout the water column can also serve as an indicator of hydrodynamic processes taking place.

Dissolved oxygen (D.O.) levels in the Hudson River estuary were determined at approximately monthly intervals at seventeen stations located between Saugerties to the north, and the apex of the New York Bight, 3.5 mi northeast of Sandy Hook to the south. The D.O. was measured by using an oxygen electrode and/or the Azide modification of the Winkler method (A.P.H.A., 1971).

Data obtained in March, July, October, November, and December of 1974 are plotted in figures 6 and 7. Data for the stations at the East River (11), Kill Van Kull (13), and lower New York Bay (15) are given separately in table 2, as these three samples come from different water masses (Chute, Rachlin, and Postmentier, 1975; Postmentier, Rachlin, and Chute, 1975), and so cannot be included in figures 6 and 7.

Inspection of figures 6 and 7 reveals some general trends which are similar from month to month. In the freshwater reaches of the river, the D.O. was higher than in the brackish or salt water regions. The D.O. declined slightly from Saugerties to Piermont, where the water was usually slightly brackish. Between Spuyten Duyvil and The Narrows the D.O. decreased, reaching a minimum in the New York Harbor, before rising again at the New York Bight station (16) in the Ambrose Channel. These trends are seen in both the surface (1.5 m below the surface) and bottom (2 m above the bottom) D.O.

In absolute terms, in the freshwater regions of the river both the surface and bottom D.O. were lowest in July when water temperature was 20°C and highest in December when it was 1°C. The difference in temperature accounted only in part for the differences in D.O. however, for in July the water was from 70 to 80% saturated with oxygen, and in December it was 85 to 93% saturated. It seems reasonable to assume that the increased water temperature in July led to increased metabolic activities of the poikilothermic heterotrophs in the river, and that this resulted in increased oxygen utilization and therefore depletion of D.O.

With the exception of July, the surface and bottom D.O. are very similar in the freshwater regions. This would suggest good vertical mixing of the waters, which may vary in depth from about 10 m at Piermont to 50 m at the World's End, but being mostly about 15 m deep. Temperature data would support this point of view.

South of Piermont the surface and bottom D.O. decreased in value, with the minimum values each month obtained in the New York Harbor between The Battery and The Narrows. Dissolved oxygen levels in the Ambrose Channel were always higher than those in the Harbor.

In New York Harbor, the lowest D.O. levels were observed in October and November, when about 4 ppm oxygen were present. At this time the water was

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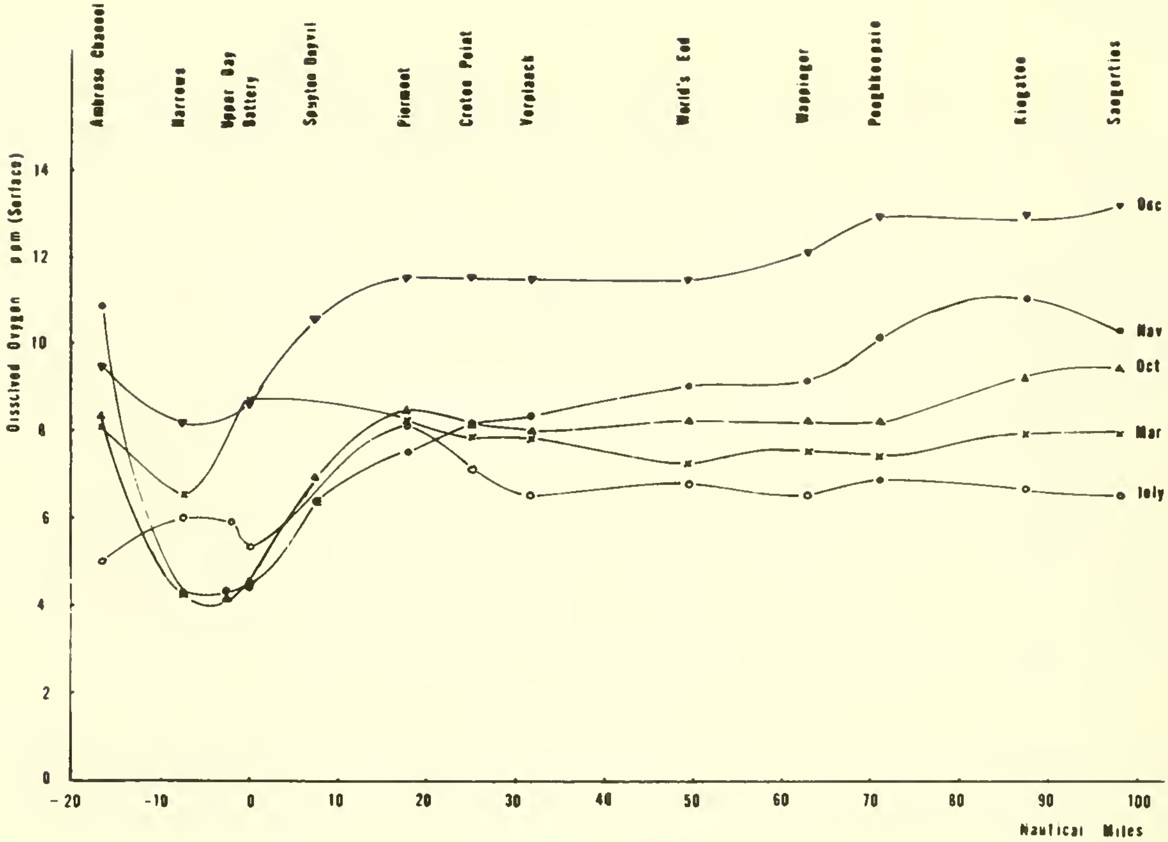

FIGURE 6. SURFACE DISSOLVED OXYGEN IN THE HUDSON ESTUARY BETWEEN SAUGERTIES AND THE AMBROSE CHANNEL FOR FIVE MONTHS IN 1974.

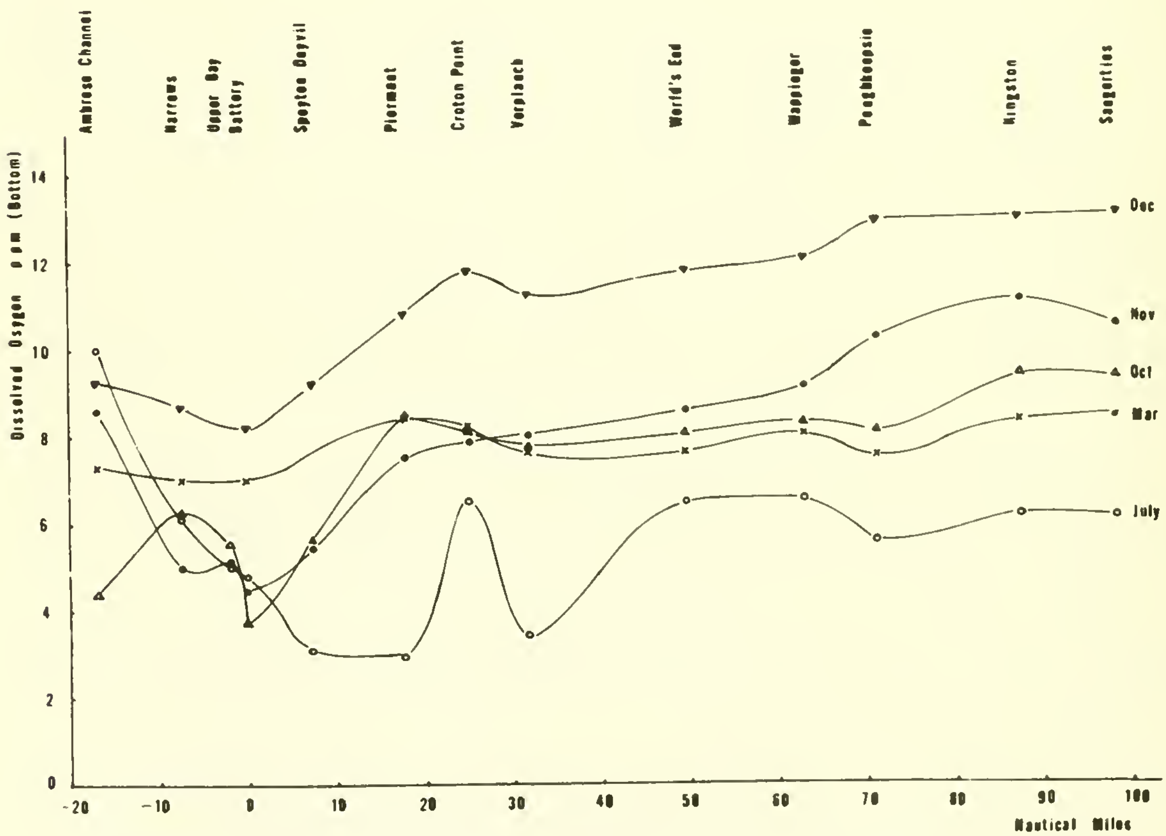

FIGURE 7. BOTTOM DISSOLVED OXYGEN IN THE HUDSON ESTUARY BETWEEN SAUGERTIES AND THE AMBROSE CHANNEL FOR FIVE MONTHS IN 1974.

about 40 to 55% saturated with oxygen. In March, July, and December, when the D.O. was higher, the percentage saturation was 55 to 70%. The depletion of oxygen found in the Harbor, as compared with the freshwater parts of the river, can not be attributed solely to the increased salinity of the harbor waters. In the absence of any studies to ascertain the processes involved (such studies are planned) it is only possible to speculate. The heavy use of the harbor by shipping of various types, and the increased sewage load carried by the water would seem to be the most obvious sources responsible for the additional oxygen depletion.

During the period under study, the D.O. of the freshwater section of the river did not fall below 6 ppm, except for two bottom samples in July. On this basis it can be concluded that there is sufficient dissolved oxygen in the Hudson River north of Piermont to provide a healthy biological environment, from the oxygen point of view, for the aquatic fauna. Even the low D.O. values found in the New York Harbor are sufficient for survival of aquatic organisms (Kinne and Kinne, 1962; Rao, 1968; Tarzwell, 1958).

It should be pointed out that, as far as the Hudson River is concerned, 1974 was not a typical year, it being much wetter than usual, thus increasing the freshwater flow rate of the river. So, while during 1974 the river had sufficient D.O. to support the life of aquatic fauna, it is quite possible that in years of lower rainfall, such favorable conditions, from the oxygen point of view might not exist.

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		Dissolved Oxygen (ppm)			
		Mar.	July	Oct.	Nov.
East River	S	12.8	5.10	3.26	3.85
	B	6.2	4.56	3.08	3.63
Kill Van Kull	S	-	5.90	3.95	4.34
	B	-	4.36	3.40	4.15
Lower Bay	S	8.1	9.13	7.12	6.43
	B	7.6	8.17	7.16	6.53

Table 2. Surface and bottom dissolved oxygen at the East River, Kill Van Kull, and Lower Bay stations for four months in 1974.

PRELIMINARY ANALYSIS OF PHYSICAL PROPERTIES OF HUDSON RIVER SEDIMENTS

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Introduction

Previous studies of the surficial sediments of the Hudson Estuary (McCrone, 1967) have provided data based on single samples taken at different locations over a period of time. This sediment study differs from the previous investigation in being part of an interdisciplinary City Institute of Marine and Atmospheric Sciences' research project aimed at both determining and interrelating the monthly parameters at the same series of stations within the Hudson Estuary.

Sample stations (Fig. 1) extend from Saugerties, New York southward to the inner New York Bight south of The Narrows. Stations were chosen to evaluate the contribution of rivers, and sources of effluents, and also to provide uniform geographic coverage over the area of study. The Hudson Estuary survey team attempted to obtain Phleger cores from each station on each monthly cruise but were often not possible due to weather, tidal current velocities, and bottom conditions.

Laboratory Analysis

The Phleger cores taken on each cruise were frozen on board ship and transferred to the laboratory in insulated containers. In the laboratory, the top 10 cm of each core was extruded, split longitudinally, described, and sampled over the entire 10 cm length. A 10 to 15 gm subsample was taken for moisture content determination and a 20 gm subsample was removed for size analysis. The 20 gm subsample was wet sieved into sand, and silt plus clay fractions. The sand fraction size distribution was determined in a set of $\frac{1}{2}$ Ø interval sonic sieves. The silt and clay fractions were dispersed in sodium hexametaphosphate with an ultrasonic probe and size analyses were made using the pipette method. Cumulative frequency curves were constructed from which the size frequency parameters of Folk and Ward (1957) were calculated.

Analysis of Data

In discussing the data presented in this paper (Table 3) it should be emphasized that only preliminary interpretations can be made on this limited data and that these interpretations will probably be considerably modified as our work progresses. Since the data presented is representative of the whole 10 cm interval, it does not accurately record the present sedimentation in the estuary but an average over the time span represented by the 10 cm of sediment accumulation.

The data for each station exhibit different degrees of variability. Many factors probably contribute to this spread of values such as small numbers of samples, vessel re-positioning errors on subsequent cruises, differential thawing of frozen cores enroute to the laboratory, and operator variation in the laboratory. In addition, natural causes in the estuary may account for

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some of this variability. Examples of these kinds of changes include artificial input resulting from channel dredging and changes in estuary currents causing erosion and deposition. Present and future research cruises are attempting to define the variability and its causes by making a series of coring traverses across the axis of the estuary.

Areal Variation in Grain Size

Mean grain sizes do not seem to show steady decreases or increases up or down the estuary but seem to be related to input from local streams. Relatively coarse sediments (fine to very fine sands) are found at station 1 (Saugerties) adjacent to the mouth of Esopus Creek and at station 2 (Kingston) adjacent to the mouth of Rondout Creek. Sediments from Poughkeepsie (stations 3A and 3B) south to Verplanck (station 6) are generally characterized by finer grain sizes (fine to very fine silt and coarse clay) with the finest material occurring at station 5 within the Hudson Gorge (World's End). Constriction of the channel at this point might be expected to result in scour and coarser sediment but the opposite seems the case. Perhaps the high cohesiveness of this material and its low bed roughness inhibits erosion once the material has been deposited from suspension.

From Verplanck (station 6) southward to the Kill Van Kull (station 13) the sediments are primarily medium to fine silts. Within this length of the estuary several interesting changes are noted. Samples taken at station 7A, which is located north of Croton Point, have finer mean grain sizes (6.5 Ø) than those on the south side of Croton Point at station 7B (5.4 Ø). This may possibly be the result of sediment input from the Croton River between the two stations. A similar difference, though not as marked, is seen on either side of the confluence of the Hudson River and the Harlem River at Spuyten Duyvil. Samples from station 9A on the north side of the confluence have a mean grain size of 6.1 Ø, whereas those from station 9B on the south side have a mean grain size of 5.93 Ø.

A distinct change in grain size occurs in the area between stations 14A (north of The Narrows) and 14B (south of The Narrows) where the average grain size is in the fine sand range in sharp contrast to the area to the north within the upper New York Bay. This may represent the effects of tidal scour due to channel constriction or the input of sand into this area by the Long Island and New Jersey littoral drift systems. Another sharp change occurs at station 15 in the lower New York Bay where the sediment is in the very fine silt to coarse clay range. Such accumulation of fines would be expected in such a wide bay away from the influx of coarser sediment.

Areal Variation in Other Physical Parameters

Standard deviation (sorting) shows no particular patterns. All of the samples are fair to poorly sorted (1.20 to 3.95 Ø). Skewness values (0.01 to 0.65) are consistently positive suggesting an excess amount of fine sediment in the size distribution. The relatively low standard deviation of the skewness values (0.07 to 0.44) relative to the standard deviation of the other parameters suggests a strongly developed trend. Perhaps this results from either greater sedimentation than erosion or from erosion and redeposition of suspended material of finer size without the development of a coarse lag between the two events.

Moisture contents (24.6 to 64.6 weight percent) show the greatest spread of values at each station of all parameters measured. This variability

may be the result of field and laboratory operations in which water was lost or moved within the core due to inadvertent defrosting of the cores in transport. A general trend of increasing water content with decreasing grain size can be seen in a plot of mean grain size versus water content.

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PRELIMINARY MINERALOGIC DATA OF HUDSON ESTUARY BOTTOM SEDIMENTS

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Clay Mineralogy

X-ray diffraction analysis of the clay fraction of Hudson Estuary bottom sediments is difficult. These sediments, for the most part, are not simple muds but organic-rich sludges. In order to obtain valid identification of clay mineral components, it is necessary to remove these organic constitutents without changing the original clay mineral structures. A number of standard techniques, alone and in combination, have been tried. As a result, preliminary data indicate that illite is the dominant clay mineral in samples obtained from Verplanck, New York and adjacent to 64th Street in New York City.

Heavy Mineralogy

Optical analysis of heavy mineral suites have been completed for two selected localities (Saugerties, station 1 and Spuyten Duyvil, station 9). In both cases data was obtained for the fine sand grade size (1/4 to 1/8 mm). Heavy mineral analysis indicate that garnet and hornblende are abundant at Saugerties, whereas at Spuyten Duyvil, there is less garnet and hornblende, and more pyroxene (notably hypersthene). The apparent variation in abundance of black opaques (mostly ilmenite) is not yet understood. Tourmaline distribution, especially the red-blue variety, may prove important relative to the definition of dispersal patterns. The study of zircons in the finer size grades may reveal significant varietal distributions.

Table 3. CORE TOP DESCRIPTIONS (upper 10 cm)

Sta. No.	Location	Wet Color Range*	Avg. Lab. Description
1	Saugerties	10Yr 3/1	Sandy silt to silty sand
2	Kingston	2.5Y 3/2-5Y 3/2 very dark grayish brown to dark olive brown	Sandy silt, thin laminae of organic material
3A	Poughkeepsie N.	5Y 3/2-5Y 4/2	Uniform silt and clay with some sand
3B	Poughkeepsie S.	2.5Y 4/2-5Y 3/1 dark grayish brown to very dark gray	Silt and clay, occasional layers of coarse material and shell fragments
4A	Wappinger Creek N.	5Y 3/2-5Y 2/2 dark olive gray to black	Uniform silt and clay, occasional small shell and wood fragments
4B	Wappinger Creek N.	5Y 3/2 dark olive gray	Uniform sandy silt and clay, occasional small pebble gravel
5	World's End	5Y 4/2-5Y 3/1 olive gray to very dark gray	Uniform silt and clay, occasional small pebble gravel
6	Verplanck	2.5 3/2-5Y 2/1 very dark grayish brown to black	Uniform silt and clay, some coarse and angular sand
7A	Croton Point N.	5Y 4/2-5Y 2/1 olive gray to black	Silt and clay, scattered angular shell fragments and shells
7B	Croton Point S.	5Y 3/1-5Y 3/2 olive gray to very dark gray	Sandy silty clay with shell fragments and pebble gravel
8	Piermont	5Y 3/1-5Y 2/1 very dark gray to black	Uniform silt and clay, scattered shell fragments, coarse sand and pebble gravel

*Munsell Color Designations

Table 3 (con't.). CORE TOP DESCRIPTIONS (upper 10 cm)

Sta. No.	Location	Wet Color Range*	Avg. Lab. Description
9A	Spuyten Duyvil N.	5Y 2/2-5Y 2/1 black	Uniform dark silt and clay, coarse sand, pebble gravel and shell in patches at top of core
9B	Spuyten Duyvil S.	5Y 5/2-5Y 3/1 olive gray to very dark gray	Sandy silt, shell fragments, medium sand, and shells concentrated near top
10	The Battery	5Y 4/2-5Y 3/1 olive gray to very dark gray	Dark silt and clay, with shell fragments and shells
11	East River	5Y 3/1 very dark gray	Uniform sandy silty clay with shell fragments more abundant in upper portion
12A	Upper Bay E.	5Y 3/1-5Y 2/1 very dark gray to black	Sandy shelly silt to shelly sand
12B	Upper Bay W.	5Y 2.5/1-5Y 2/1 black	Oily and organic silt and clay, shell fragments
13	Kill Van Kull	5Y 2.5/1-5Y 2/1 black	Uniform silt and clay, oily and organic, shell fragments
14A	The Narrows N.	2.5Y 3/2-5.5Y N 3/0 very dark gray brown to dark gray	Sandy silt, worm tubes running vertically through core
14B	The Narrows S.	2.5Y N 2/0 black	Uniform silt and clay, oily sand and shell fragments near top
15	Lower Bay	5Y 2/1	Plastic silt and clay, oily and organic, trace of fine sand

*Munsell Color Designations

FORAMINIFERAL ASSEMBLAGES IN THE HUDSON ESTUARY

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Within the study area of the Hudson Estuary survey of the City Institute of Marine and Atmospheric Sciences, foraminifers have been found south of Verplanck, New York (Fig. 8). Samples for foraminiferal analysis have been collected at stations 6 to 10, 12, 14 to 16 (Fig. 1). Most stations were occupied monthly from December, 1973 to December, 1974 to determine if temporal changes occurred with respect to foraminiferal distributions. Samples were taken using a Phleger gravity corer from which the upper cm of sediment (approximately 25 cm^3) was removed for foraminiferal analysis. All sediment samples collected were treated with rose bengal stain dissolved in methanol. The stain solution served as a preservative and as a means to distinguish living (stained) from dead (unstained) foraminifers. Environmental data (i.e., salinity, temperature, dissolved oxygen, turbidity) were recorded for each sampling station. In the laboratory, samples were washed onto sieves with mesh openings of 125 and 63 microns. Foraminifers were then separated from the bulk of the sediment trapped on the 125 micron screening by the carbon tetrachloride flotation method. A binocular microscope was used to identify the foraminifers present in each sample.

Foraminifers representing 15 genera and 20 species have been identified (Table 4). Total numbers of foraminifers per sample ranged from zero to over 2500. Specimen representing live foraminifers were not found in all samples or representative of every species identified.

Three distinct assemblages of foraminifers have been identified to date (Fig. 8, Table 4). The assemblages are as follows:

1. Ammobaculites assemblage. The assemblage is dominated by agglutinized (arenaceous) foraminifers. Ammobaculites and Ammomarginulina are the dominant forms present followed by lesser amounts of Trochammina and Miliammina. In addition, an occasional specimen of Elphidium and Ammonia have been found. This assemblage is found in an area from station 6 (Verplanck) on the north to station 9 (Spuyten Duyvil) on the south (Figs. 1, 8). The salinity for this section of the estuary during the period of study averaged about $6^{o}/oo$ and ranged from 0.4 to $15^{o}/oo$. The northernmost limit of the Ammobaculites assemblage is indicative of the maximum penetration of foraminifers into the Hudson Estuary. In addition, this uppermost limit can serve as a marker for the boundary between oligohaline and mesohaline waters at a salinity of $5^{o}/oo$.

2. Elphidium - Ammonia assemblage. The Elphidium - Ammonia assemblage is characterized by species of Elphidium, especially E. clavatum - incertum complex, and lesser amounts of A. beccarii. In addition, agglutinized foraminifers such as Ammobaculites have also been found in this assemblage. The Elphidium - Ammonia assemblage has been identified at stations 9 (Sputen Duyvil), 10 (The Battery), 12 (Upper New York Bay), and 15 (Lower New York Bay) in waters with bottom salinities ranging from 11.2 to $26.1^{o}/oo$, and averaging $20^{o}/oo$ (Figs. 1, 8). This assemblage occupies a transitional position between the lower salinity

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indicative Ammobaculites assemblage and the higher salinity, more open water, Elphidium - Quinqueloculina assemblage. It overlaps the former on the north and the latter on the south (Fig. 8).

3. Elphidium - Quinqueloculina assemblage. Elphidium clavatum - incertum complex and Quinqueloculina seminula are indicative of the third foraminiferal assemblage. Species assignable to the E. clavatum - incertum complex are found in great abundance. Specimen of Buccella frigida, Ammonia beccarii, Discorbis squamata, and Cibicides lobatulus were also found. Relatively small numbers of agglutinized foraminifers have also been found in this assemblage along with individual species characteristic of the waters of the inner New York Bight and western Long Island Sound. This assemblage is found at stations 14 (The Narrows) and 16 (New York Bight). Bottom water salinities during the study period in the area of occurrence of this assemblage averaged 29.0°/oo and ranged from 25.7 to 32.4°/oo.

The examination of foraminiferal samples on a monthly basis indicates that the distribution of the assemblages described above is apparently controlled by the salinity of the estuary. During the high salinity months of July to November, the assemblages display their most northerly distribution in the estuary. The boundaries between the assemblages, along with the maximum limit of foramifers, shifts as much as 20 mi to the south during December to June, the low salinity phase in the estuary.

Table 4. FORAMINIFERAL ASSEMBLAGES. Species listed in descending abundance.

AMMOBACULITES ASSEMBLAGE	ELPHIDIUM - AMMONIA ASSEMBLAGE
Ammobaculites sp.	Elphidium clavatum - incertum complex
Ammomarginulina sp.	Ammonia beccarii
Trochammina sp.	E. subarcticum
Miliammina fusca	Ammobaculites sp.
Elphidium clavatum - incertum complex	Ammomarginulina sp.
Ammonia beccarii	Buccella frigida
	Quinqueloculina seminula

ELPHIDIUM - QUINQUELOCULINA ASSEMBLAGE

Elphidium clavatum - incertum complex	Q. seminula var. jugosa
Ammonia beccarii	Q. subrotundra
Quinqueloculina seminula	Cibicides lobatulus
E. subarcticum	Triloculina trihedra
Buccella frigida	Nonionella auricola
Discorbis squamata	Discorbis columbiensis
Nonionella atlantica	Trochammina sp.
Pseudopolymorphina novangliae	Ammobaculites sp.

DIATOM DISTRIBUTION IN THE HUDSON ESTUARY

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Diatoms are important members of most aquatic communities and often the major primary producers. Weaver (1970), in a year long study of phytoplankton in the Hudson River at Indian Point, found species of the diatom genus Melosira to be the dominant phytoplankter, with the diatom Asterionella also occurring in high numbers. Whereas Weaver used plankton nets with a mesh size of 76 microns, this study has invovled the analysis of diatom assemblages in bottom sediments. Thus, live cells and empty frustules are being examined. In addition, diatoms as small as 5 microns are being studied.

The Hudson River and Estuary system has an abundant and diverse diatom flora. In a count of 350 individuals it is not uncommon to have as many as 55 to 60 species, although most of them occur in low numbers and only a few reach frequencies as high as 10%. In sediment samples Stephanodiscus astrea (sensu lato) is the dominant form from Poughkeepsie, New York (station 3) to Upper New York Bay (station 12). Weaver (1970) did not find the planktonic Stephanodiscus to be an abundant genus. It is generally much smaller than 76 microns and thus was not trapped in her plankton tows. Melosira sulcata is dominant in the Upper New York Bay (station 12), and Anorthoneis hyalina is dominant at the New York Bight station (16A).

On the basis of selected diatom distributions (Fig. 9), the study area can be divided into four zones as follows:

Zone 1: marine - high salinity brackish: inner New York Bight to The Battery (samples 16A to 10).
Zone 2: brackish; Spuyten Duyvil to Tarrytown (samples 9A to T10C).
Zone 3: brackish - very low salinity; Tarrytown to West Point (samples T12C to 5).
Zone 4: freshwater: Newburgh to Saugerties (samples 4B to 1).

These divisions are based primarily on the distribution of marine and brackish water species as the freshwater species tend to have their distribution extended by downstream transport of frustules. This is especially evident in the distribution of the diatom species Stephanodiscus astrea (Fig. 9).

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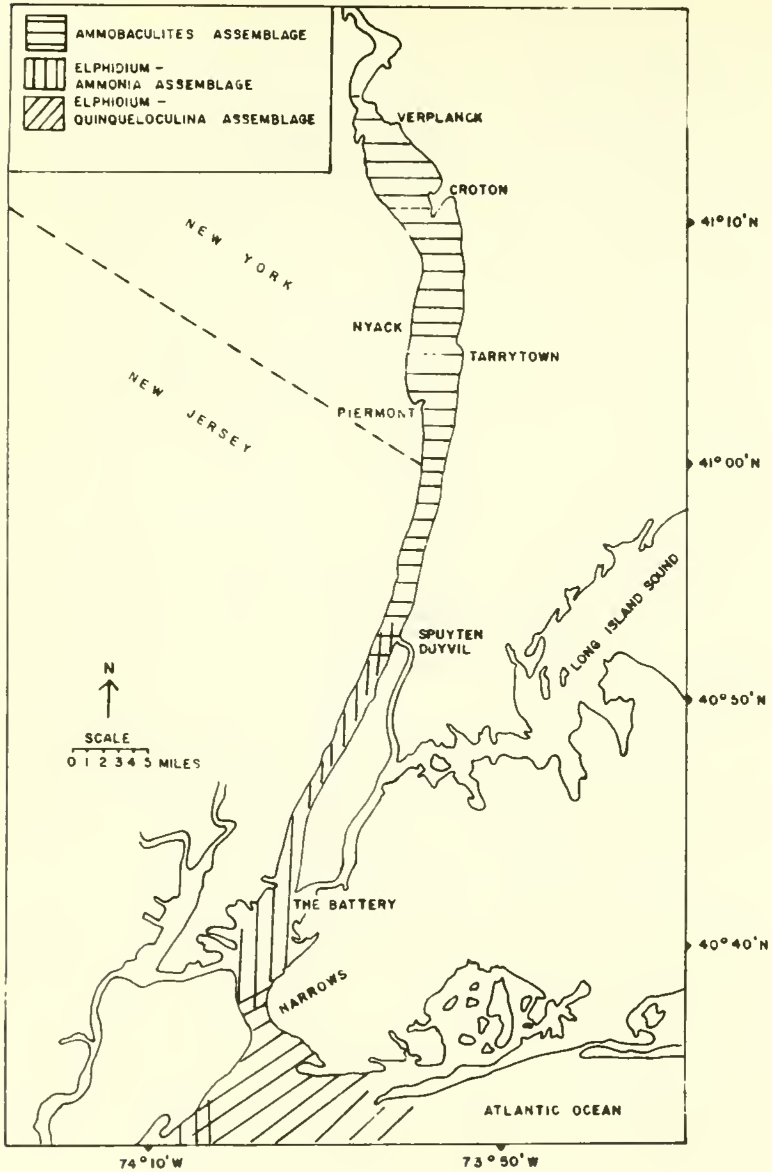

FIGURE 8. DISTRIBUTION OF FORAMINIFERAL ASSEMBLAGES IN THE HUDSON ESTUARY.

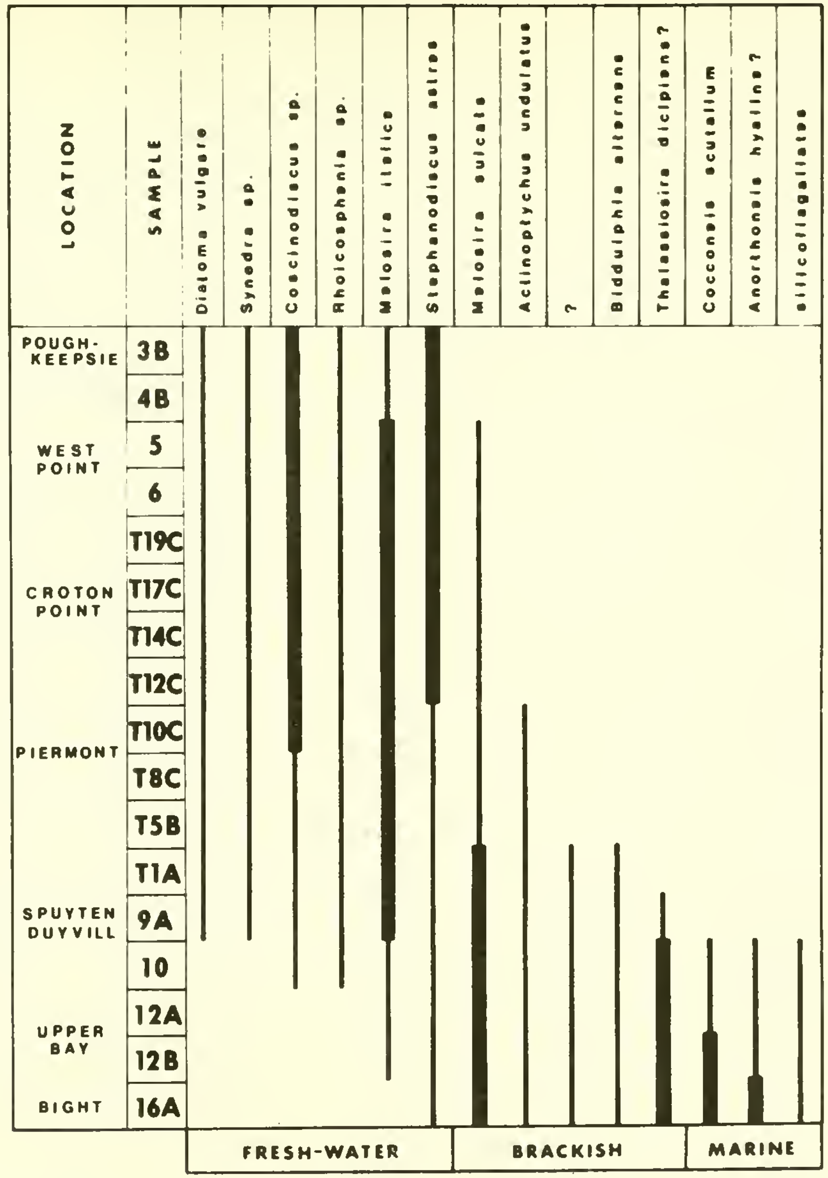

FIGURE 9. DISTRIBUTION OF IMPORTANT DIATOM SPECIES. The width of each line representing specie's range indicates the area in which that species occurs most abundantly.

DISTRIBUTION OF SHELLED MACROINVERTEBRATE BENTHOS IN THE HUDSON ESTUARY

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Shelled benthos (primarily molluscs, bryozoans, and barnacles) have been sieved from one liter bottom grab samples. Samples have been taken on a random monthly basis for one year at the 16 stations of the City Institute of Marine and Atmospheric Sciences' Hudson Estuary survey, and are located in or near the main ship channel from the inner New York Bight (station 16) to Saugerties, New York (station 1). The number of live individuals, disarticulated valves, fragments, and preliminary size measurements have been recorded. Three assemblages, characterized by nominate bivalve genera and consisting largely of bivalves are proposed:

1. Elliptio assemblage: an essentially freshwater fauna present from Peekskill to Saugerties.
2. Mytilopsis assemblage: occupying the area between Yonkers and northern Haverstraw Bay, showing low species diversity in bottom salinities which may vary as much as 15^{o}/oo in the southern Tappan Zee. Salinities over this entire reach range from 1 to 15^{o}/oo.
3. Mulina - Mya assemblage: essentially marine in generic composition and diversity, present as far north as Riverdale in salinities which may drop as low as 10^{o}/oo.

A typical monthly sample of these assemblages is shown in figure 10 and compared to other proposed estuary classifications in figure 11.

Bottom salinity and temperature appear to be the controlling factors for these assemblages. Dissolved oxygen is judged adequate for molluscs at all stations over the sampling area. Bottom sediments in the study are have not been fully analyzed. There does appear to be little difference in the genera at stations 10, 12, and 13 in spite of major differences in sediment.

A finer grid lateral to the main channel is being sampled at present to fully clarify the local factors controlling the distribution of shelled macroinvertebrate benthos.

Several questions may now be raised on the basis of this new information. First, do local distruptions in this assemblage pattern, such as the very scanty faunas in the south Yonkers - Riverdale area and the demise of the shell fish industry in the Tappan Zee, represent human interference or natural changes in the Estuary? Secondly, can the distribution and individual growth characteristics of bivalves be used as part of a natural monitoring system in the Estuary?

Contribution No. 55 of the City Institute of Marine and Atmospheric Sciences.

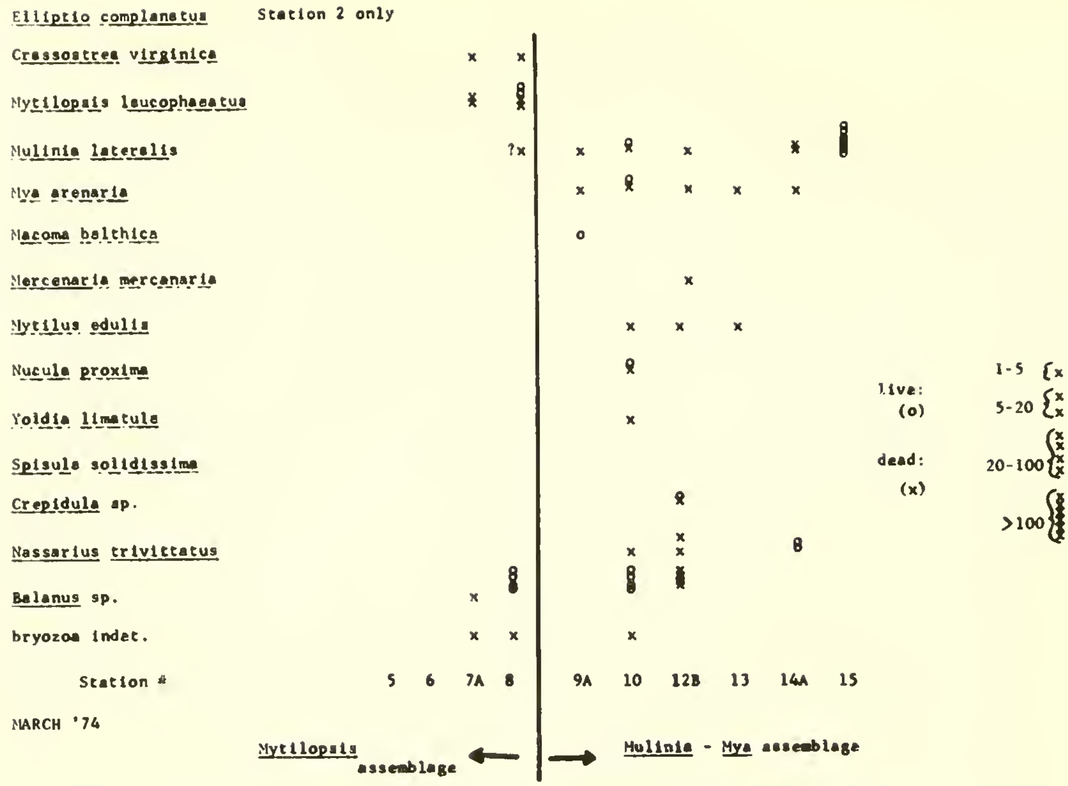

FIGURE 10. SELECTED TALLY OF HUDSON RIVER BENTHIC SPECIES. Numbers refer to survey stations. 5 is to the north.

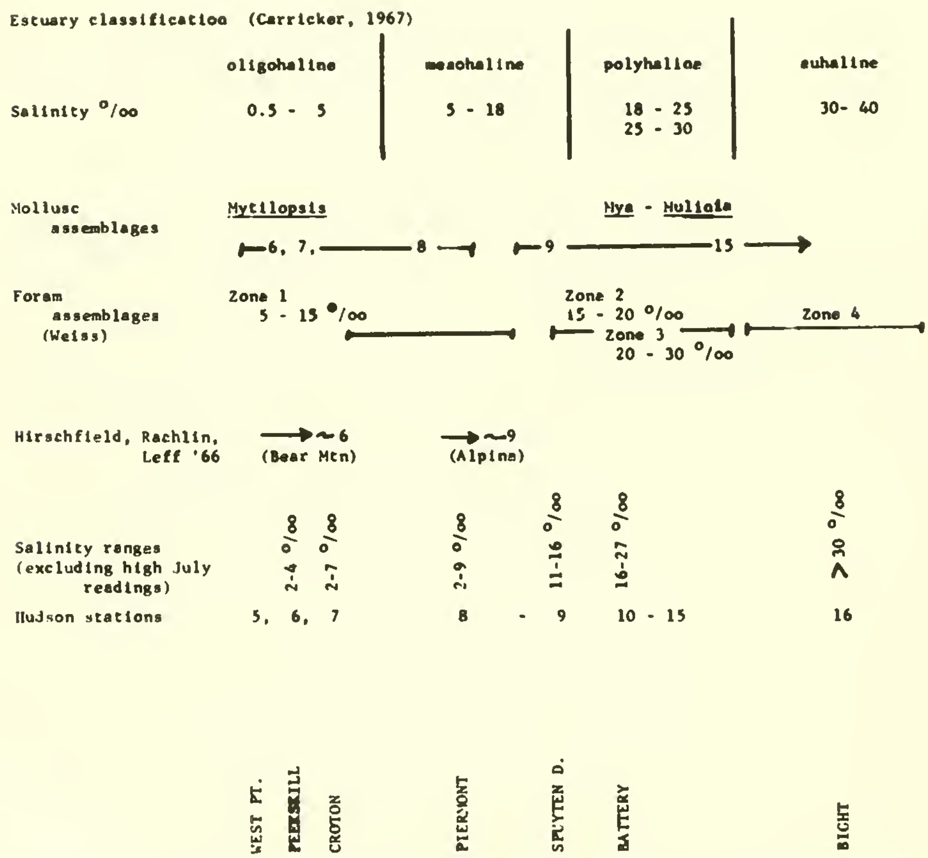

FIGURE 11. COMPARISON OF MULLUSCAN ASSEMBLAGE DISTRIBUTIONS TO FORAMINIFERAL DISTRIBUTIONS OF WEISS AND STANDARD ESTUARY CLASSIFICATION OF CARRICKER.

STRUCTURAL AND STRATIGRAPHIC CHRONOLOGY OF THE TACONIDE AND ACADIAN POLYDEFORMATIONAL BELT OF THE CENTRAL TACONICS OF NEW YORK STATE AND MASSACHUSETTS

Nicholas M. Ratcliffe[1], John M. Bird[2], and Beshid Bahrami[1]

Introduction

On this trip we will examine selected features of the Taconic rocks that have a bearing on deciphering the complex depositional and tectonic events that have affected these rocks. The traverse at $42^{\circ}15$ N. latitude extends from the low Taconics near the Hudson River, east to the high Taconics of Massachusetts. The discussion, rather than being complete, is selective and stresses new evidence not previously treated. For a general discussion, see Trip B-1. Some controversial questions will be raised; we hope to stimulate new interest in Taconic geology by pointing out some of the major unsolved problems that remain to be studied in this and presumably other parts of the Taconic allochthon.

In the past decade major advances in our conceptual knowledge of the Taconic geology have been made largely by the efforts of E-an Zen (1961, 1967, 1972), Bird and Dewey (1970), and Bird and Rasetti (1968). As a result of these studies, a unified picture has evolved that is elegantly simple but at the same time incredibly comprehensive. However, some of the Taconic rocks are exceedingly complex, and many unresolved problems remain to be studied.

For a recent summary of the geology of the Giddings Brook slice at this latitude see Bird and Dewey, Trip B-1.

Zen (1967) has proposed that the allochthonous rocks of the Taconics belong to six or seven discrete structural slices (fig. 1) that overlap eastward so that the highest structural level, the Dorset Mountain slice, in western Massachusetts, known as the Everett slice (Ratcliffe, 1969), crops out at the east edge of the allochthon. Rocks of the Everett slice constitute the high Taconic sequence at this latitude and are presumed to have been emplaced last.

The low Taconics here are represented by rocks of the Giddings Brook, Chatham, and Rensselaer Plateau slices according to Zen (1967). The distinction between high and Taconic is based in part on topographic expression, relative structural position, and stratigraphic considerations and implies no one specific tectonic or stratigraphic attribute. The terminology is imprecise and probably has outlived its usefulness.

Zen further proposed that the stratigraphic range of the individual slices is greatest in the lowest slices and most abbreviated in higher

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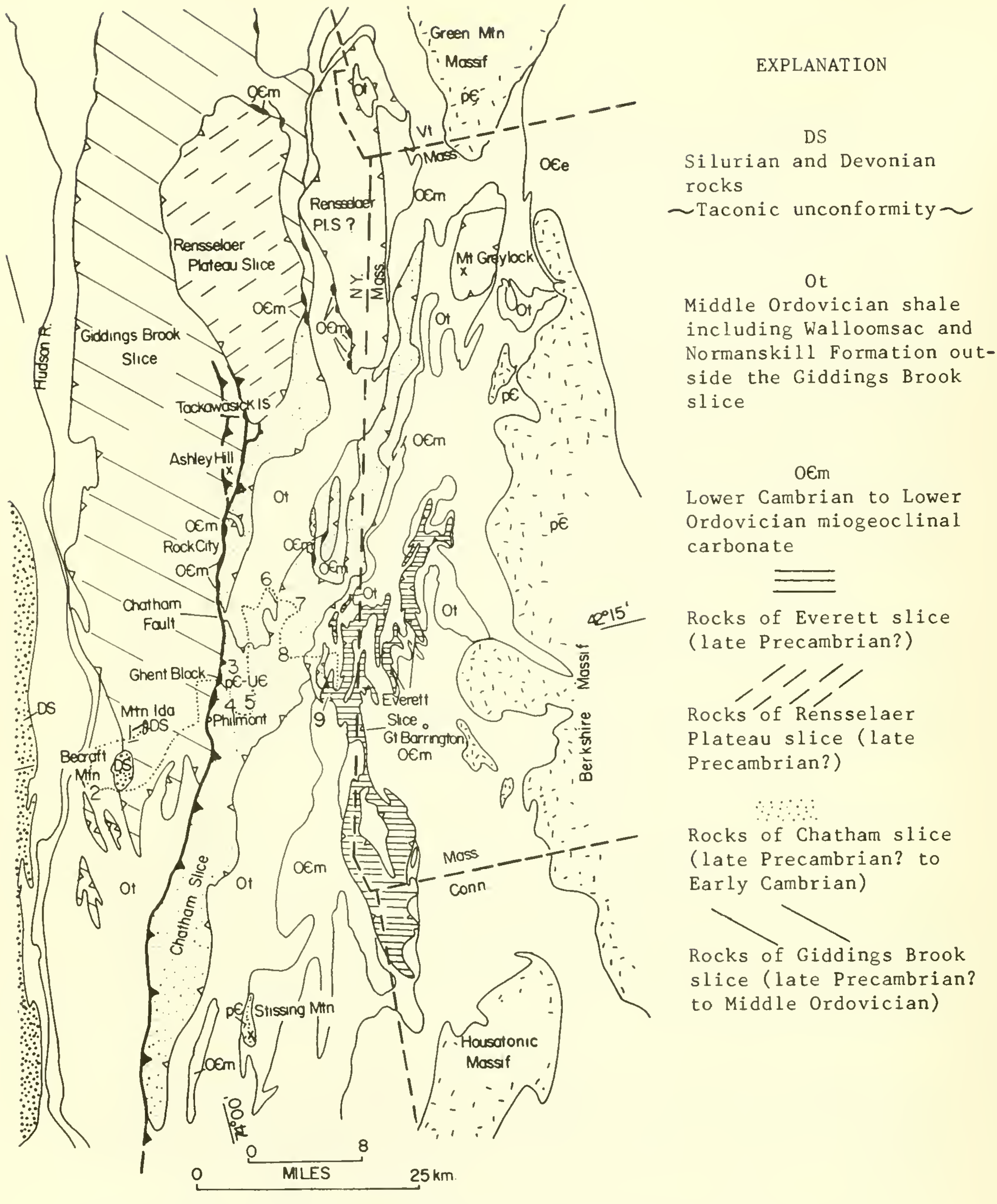


Figure 1. Regional geologic map showing slices of the Taconic allochthon modified from Zen (1967), based on data in Ratcliffe (1974a), Ratcliffe and Bahrami (in press), and Potter (1972). The Chatham fault is shown with solid triangles; extensions north of Rock City and south of Philmont are conjectural.

slices, which contain rocks largely of inferred late Precambrian age (Zen, 1967). The lowest and westernmost slices, the Giddings Brook and Sunset Lake (in Vermont), were emplaced by gravity gliding in the Middle Ordovician, contemporaneously with wildflysch-like (Forbes Hill Conglomerate) material that contains fossiliferous and nonfossiliferous fragments of the allochthon itself. Graptolites of Zone 13 (Berry, 1962, p. 715) in the matrix of the wildflysch-like conglomerate that underlies the Giddings Brook (East Petersburg slice of Potter, 1972) and Sunset Lake slices date the time of submarine emplacement (Zen, 1967; Bird, 1969). Graptolites of Zone 12 (Berry, 1962) have been collected from the Walloomsac Formation which underlies wildflysch-like conglomerate at the eastern (trailing) edge of the Giddings Brook slice (North Petersburg slice) at Whipstock Hill (Potter, 1972). This suggests that the Giddings Brook slice was emplaced during the timespan represented by Zones 12 and 13, although the lack of fossils in the matrix at Whipstock Hill precludes proof of this point.

The Chatham slice overrides the Giddings Brook slice along the Chatham fault of Craddock (1957) (fig. 2). The fault zone contains slivers of carbonate and other rocks (see discussion, Stop 3) that were thought to have been plucked from the autochthon during emplacement in Sherman Fall time (Zen, 1967, p. 34). To the east, the Chatham slice is overlain by the Everett slice at the sole of which are distinctive tectonic breccias that consist of complex mixtures of fragments of all the shelf sequence carbonates, and Walloomsac and Everett, lithologies concentrated along the soles of imbricate slices (Zen and Ratcliffe, 1968; Ratcliffe, 1969, 1974a) (Stop 9).

Chatham slice and the Chatham fault

The rocks of the Chatham slice were studied previously by Craddock (1957) and Weaver (1957), who did not map detailed stratigraphy within the slice. Thus Zen in his 1967 compilation had only limited data available bearing on Chatham slice stratigraphy.

The results of recent detailed mapping in two quadrangles spanning the width of the Chatham slice are shown in Figure 2. Rocks of the Chatham slice resemble closely gray-green and purple slate (Mettawee), Rensselaer graywacke, and other rocks of the Nassau Formation (Bird, 1962a) in the Giddings Brook and Rensselaer Plateau slices (Table 1). The Chatham slice sedimentary rocks (Nassau) probably also are pre-Olenellus in age. Distinctive but sporatically developed diabasic basalts, pillow lavas, and pyroclastic volcanic rocks are spatially associated with the base of the Rensselaer facies in all three slices (Balk, 1953; Potter, 1972; Ratcliffe, 1974a).

Massive quartzites similar to the Zion Hill (Zen, 1961) and Curtis Mountain quartzites (Fisher, 1962) crop out in the Chatham slice within the areas denoted by Ꞓnsq on Figure 2. However, the quartzites mapped in the Chatham slice in both the State Line and Chatham quadrangles underlie

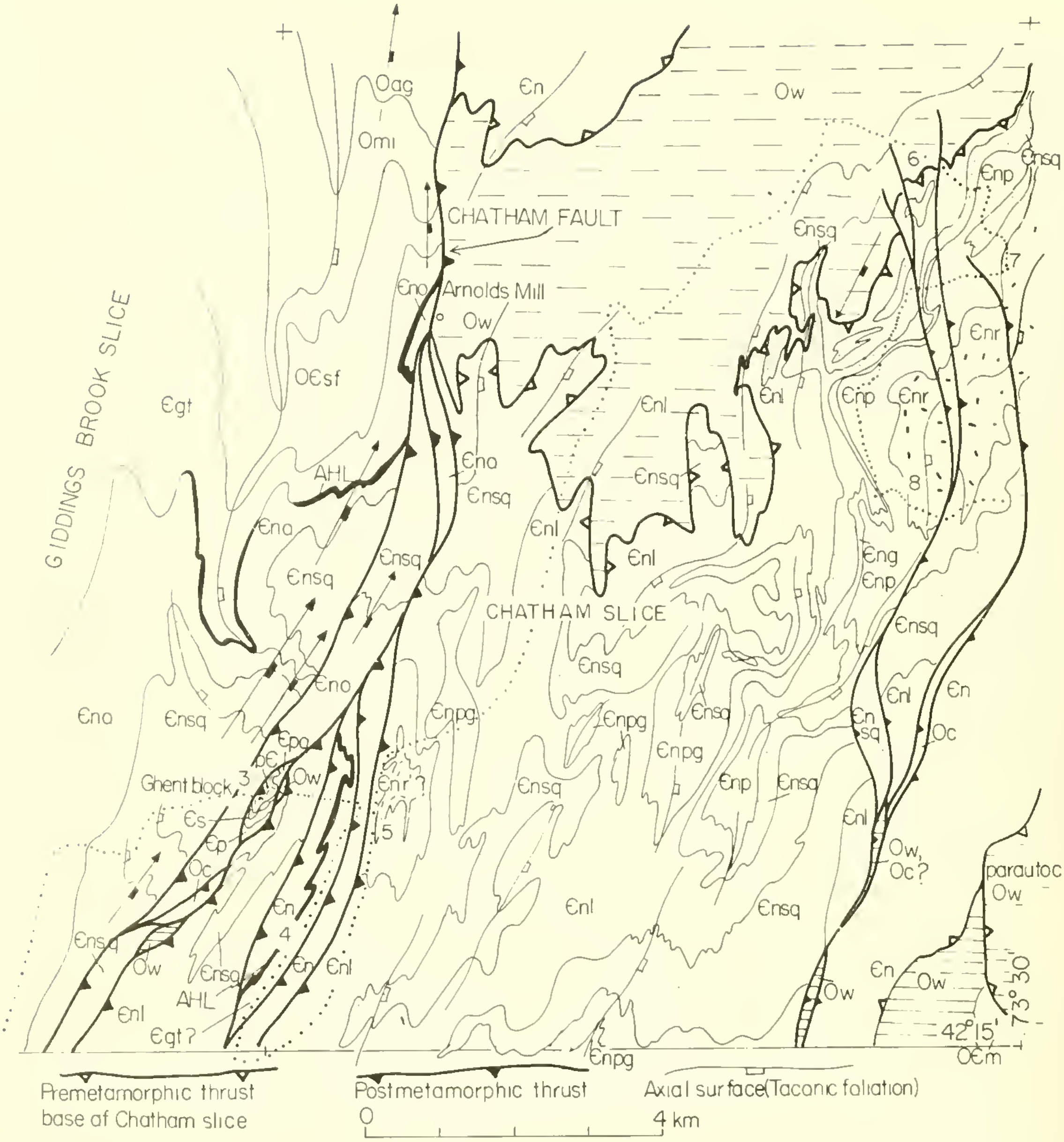
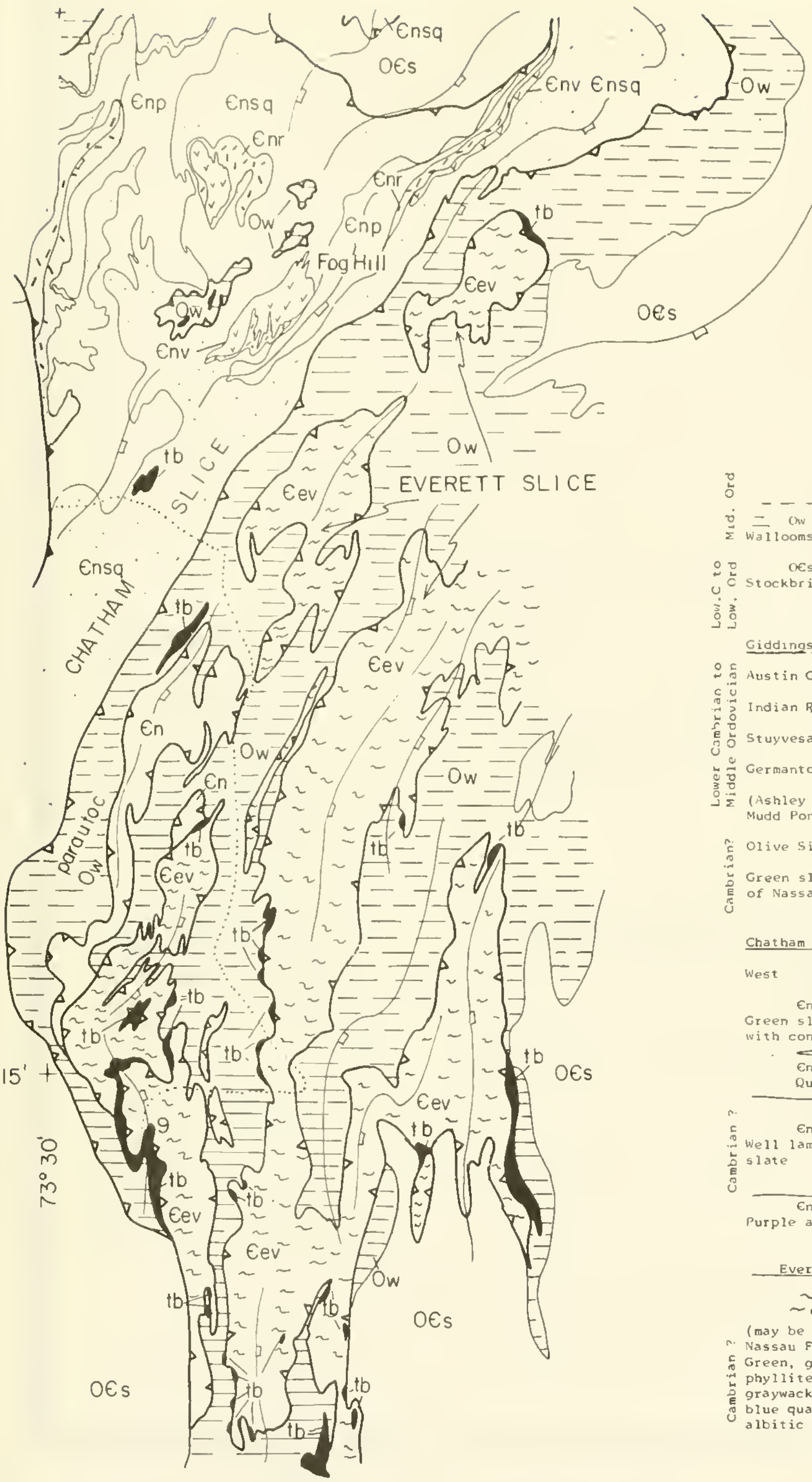


Figure 2. Generalized geologic map of the contact relationships between the Giddings Brook, Chatham, and Everett slices of the Taconic allochthon in the Stottville and Chatham, New York, and State Line and Egremont, Massachusetts and New York, quadrangles. Saw teeth indicate thrust fault, teeth on upper plate; open triangle indicates a premetamorphic fault; solid teeth indicate postmetamorphic (post-Taconic) fault. Maps join at 73°30' W. longitude. Field trip route, Stops 3 to 9, indicated by dotted line.



EXPLANATION

Bedrock of Autochthon

Mid. Ord

Ow
Walloomsac Formation

tb
tectonic breccia at sole of Everett slice composed of blocks of OЄs, Ow, and Єev intermixed

Low.C to Low. Ord

OЄs
Stockbridge Formation

Bedrock of allochthon

Giddings Brook slice

Lower Cambrian to Middle Ordovician

Oag
Austin Glen Graywacke
Omi
Indian River-Mount Merino
OЄsf
Stuyvesant Falls Formation
Єgt
Germantown Formation
AHL
(Ashley Hill Limestone conglomerate, Mudd Pond Quartzite)

Cambrian?

Єno
Olive Siltstone of Nassau Formation
Єnsq
Green slate, quartzite and siltstone of Nassau Formation

Chatham slice (shows schematic west to east facies relationships)

West — East

Cambrian ?

Єnsq
Green slate, quartzite siltstone with conglomerate

Єnr? Curtis Mountain Quartzite?

Єnl
Well laminated green, gray slate

Єnpg
Purple and green slate

NASSAU FORMATION

Єnr
Rensselaer graywacke

Єnv
basalt flows, andesite?

Єnsq
Green slate siltstone quartzite

Єnp
Purple slate

Everett slice

Cambrian ?

Єev
(may be equivalent to lower part of Nassau Formation)
Green, gray, purple and green phyllite with minor beds of graywacke and greenish quartzite; blue quartz pebble conglomerate, albitic phyllite

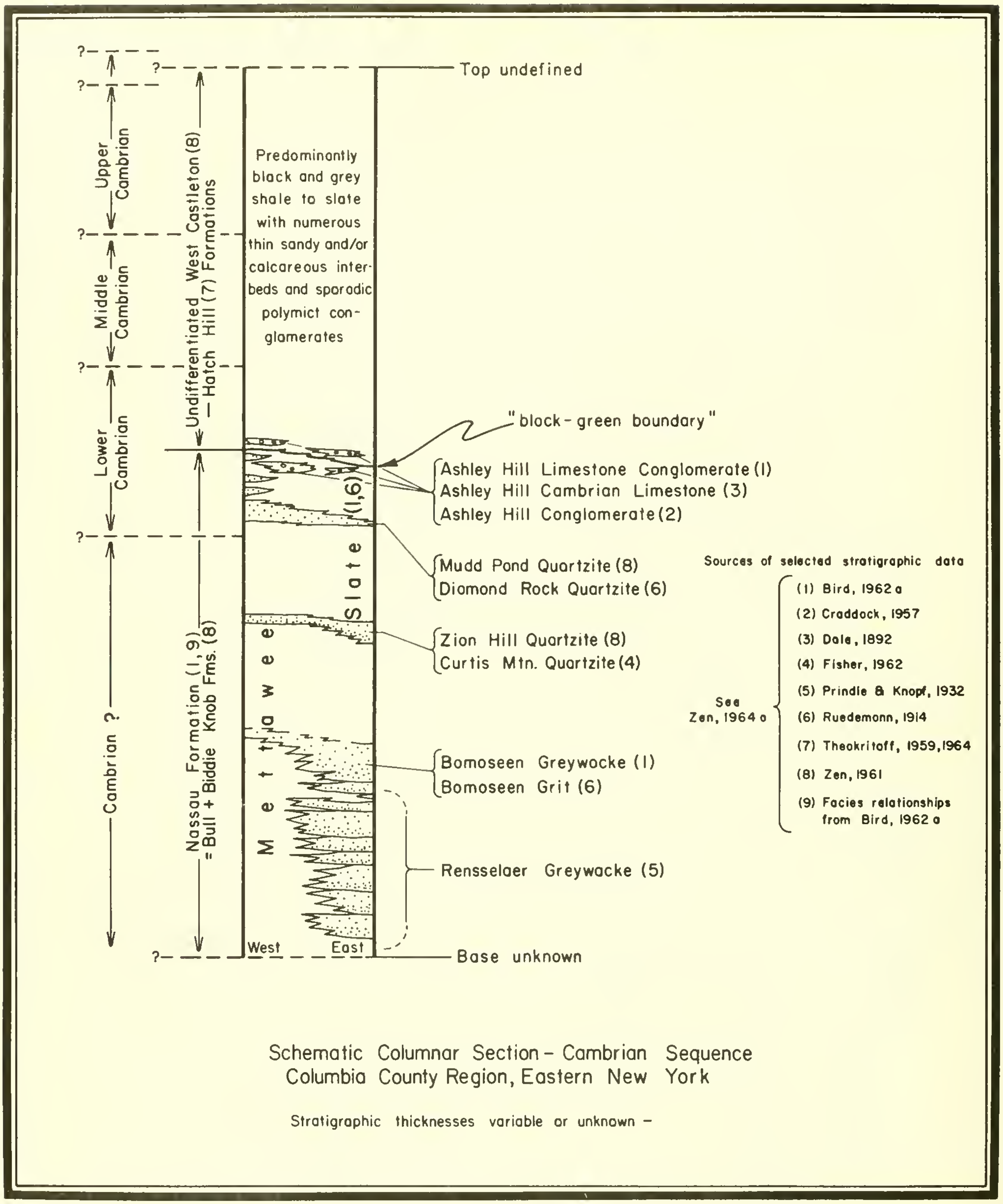


Table 1. Stratigraphic sequence of Cambrian rocks in the Giddings Brook and Rensselaer Plateau slices of Columbia County, New York. Reproduced from Bird and Rasetti (1968). Stratigraphy of similar rocks of the Chatham slice is shown in the explanation of Figure 2.

the Rensselaer-type graywackes. One of these quartzites with a polymict basal conglomerate (Stop 5) contains angular fragments of basaltic or andesitic scoria, suggesting that the relatively thin subgraywackes and quartzites exposed in the western part of the Chatham slice may be tongues of Rensselaer-like material that extended westward into the sedimentary basin.

Importantly, the Rensselaer-like graywacke of the Chatham slice in the Austerlitz outlier and in the State Line quadrangle overlies a considerable thickness (2,000 to 2,500 feet) of purple and green slate, siltstone, and laminated green slate typical of the Nassau elsewhere. However, Rensselaer graywacke of the Giddings Brook and Rensselaer Plateau slices appear at or near the base of the preserved stratigraphic succession (Table 1). The stratigraphic position of the Rensselaer within the original (as opposed to the allochthonous) sequence is really moot, because the original sequence is nowhere preserved intact, and we do not know at present if the Chatham relationships are the rule rather than the exception.

Rocks probably as young as the Ashley Hill Limestone and the West Castleton-Hatch Hill sequence have recently been discovered (Stop 4) by Bahrami within the area of the Chatham slice as shown by Zen (1967), who used Craddock's (1957) location of the Chatham fault, the westernmost fault on Figure 2. At present we believe this sequence is a large fault sliver rather than a proven part of the Chatham slice stratigraphy. Figure 2 shows that stratigraphic units common to each slice are contiguous along the Chatham fault zone south of Chatham, and Nassau stratigraphy in the two slices overlaps.

The internal structure of the Chatham slice is complex and not fully understood at present. Although no major recumbent folds are recognized, a series of F_1 folds older than the slaty cleavage is present (fig. 2). From Arnolds Mills eastward, the rocks of the Chatham slice clearly overlie the Middle Ordovician Walloomsac, based on the detailed analysis of minor structures by Bahrami. Locally a wildflysch-like conglomerate is present within meters of the contact (Stop 6). The regional Taconic slaty cleavage crosscuts the thrust, and the contact is folded into northwestward overturned folds. The allochthonous nature of the Chatham slice has been demonstrated by geometric relationships (Ratcliffe, 1969, 1974a; Ratcliffe and Bahrami, in press), and the field relationships are compatible with emplacement in the Middle Ordovician. The wildflysch-like rock seen at Stop 6 supports this conclusion.

Detailed mapping in the Chatham and Stottville quadrangles has shown that the Chatham fault (contact between the Giddings Brook and Chatham slices) is a late tectonic feature and that the original Ordovician boundary between the two slices may no longer exist intact. The Chatham fault is a major post-Taconic (post-slaty cleavage) thrust that encorporates fragments of shelf carbonates, pieces of the Chatham and Giddings Brook slices, and a block (Ghent block) of Grenville gneiss with attached rocks of the shelf sequence (Stop 3). Because of the demonstrable late origin of the Chatham fault, and because of the overlap in Nassau stratigraphy, no direct

evidence requiring that the Giddings Brook and Chatham slices be separate slices is known. Ratcliffe and Bahrami (in press) suggest that they were continuous prior to imbrication in the Chatham fault.

Because of the geometric relationships cited above and the extensive overlap in Nassau stratigraphy among all three slices, it does not seem likely that the Giddings Brook and Chatham slice rocks were deposited directly above the rocks of the Rensselaer Plateau slice (Zen, 1967, p. 67) in the Taconic depositional basin. The rocks of the Rensselaer Plateau slice that now overlie the Giddings Brook slice (Potter, 1972) could have been deposited either east or west of the rocks of the combined Chatham-Giddings Brook slice. Combined the Chatham and Giddings Brook slices are 35 km wide at 42° N. The amount of tectonic shortening is unknown, but an original depositional site 50 km wide somewhere to the east (present compass direction as opposed to the south for late Precambrian time) appears to be the minimum distance necessary to accommodate these rocks. Ratcliffe (1969, 1974a) has shown that the rocks of the Chatham slice extend northward along the New York-Massachusetts line to connect with the belt of rocks shown as Dorset Mountain slice by Zen (1967, fig. 2) west of Pittsfield, Mass. This change is encorporated in Figure 1.

Everett slice

Rocks of the Everett Formation that form the high Taconic Everett slice at this latitude are greenish-gray, green, and locally purplish slate with relatively minor amounts of interbedded Rensselaer-like graywacke. In general the Everett resembles rock of the lower Nassau Formation when the effect of increased metamorphic grade is considered. Zen and Hartshorn (1966), Zen and Ratcliffe (1968), and Ratcliffe (1969a, 1974a, 1974b) consider the Everett rocks to be as old or older than rocks of the western slices. No fossils have ever been found within rocks of the Everett slice, and are not likely to be, so that the age problem may never be completely resolved. The Everett slice is about 12 km wide and probably originated from a depositional site at least this wide. Internal structure within the Everett slice, however, is poorly known, owing to the lack of coherent stratigraphy; the possibility of stacked slices of material that all rooted from the same zone could reduce this 12 km figure.

The contact relationships of the Everett and Chatham slices are complicated because the leading edge of the Everett slice is a zone of intense imbrication involving both allochthonous and autochthonous rocks. A belt of parautochthonous Walloomsac everywhere separates the two slices (fig. 2). Locally slivers several km long of purple and green slates typical of Chatham slice rocks are found encorporated in the parautochthonous belt of Walloomsac. In addition, at least two imbricate slices of Everett rocks are found above the Walloomsac sliver and above the slivers of Chatham slice rocks (Ratcliffe, 1974a).

The contact of parautochthonous Walloomsac on the Chatham slice and between the Everett and all other rocks is marked locally by an intensely

developed tectonic breccia composed of inclusions of Stockbridge Formation. These breccias mark tectonic movement zones that differ from conventional fault zones in one important aspect. The carbonate clasts in the highly imbricated slate matrix are exotic blocks not derived from the present hanging wall or foot wall but from the autochthonous Stockbridge belt, and thus are considered tectonic inclusions transported within the movement zone from some site to the east. The tectonic breccia is evidence for a thrust beneath the Everett slice, which is independent of the regional stratigraphic arguments (Zen and Ratcliffe, 1966). These breccias have been mapped throughout southwestern Massachusetts (Zen and Ratcliffe, 1968; Ratcliffe, 1974a, 1974b) and are found in east and west dipping contacts as well as along the nose of plunging folds of the thrust contacts. The emplacement of the breccias predated the first regional metamorphism and the penetrative foliation that crosscut the contact of the thrust slices with the autochthon. Emplacement of the Everett slice resulted in brittle deformation (plucking) of the carbonate rocks, indicating that the carbonates were lithified at the time of thrusting. Similar brittle deformation of the pelitic rocks is not recognized, although an abnormally strong phyllitic foliation has been noted by Zen (1969) (see Stop 9, this trip) immediately adjacent to the carbonate slivers. It is commonly suggested, therefore, that the Everett slice was the result of hard rock thrusting because of these breccias. However, confirmation of the hard versus soft character of the allochthonous rocks must come from evidence in the allochthonous rocks. Thus far, clear evidence for brecciation is lacking, and the Everett slice need not have been completely indurated at the time of emplacement.

The age of emplacement of the Everett slice is unknown, but based on geometric relationships its final movements postdated emplacement of the Chatham slice in Middle Ordovician and predated formation of the regional slaty cleavage that probably is Late Ordovician in age (see Table 2).

A higher slice of Taconic-like rocks crops out near Great Barrington on June Mountain (Ratcliffe, 1974c) and on Canaan Mountain in the Ashley Falls quadrangle (Ratcliffe and Burger, 1975; Harwood, U.S.G.S. unpub. data, see Trip B-2). Rocks of this slice have a post (M_1) metamorphic emplacement fabric (Table 2), and Ratcliffe (1974c) suggests that these rocks are part of an extensive sheet of Taconic-like rocks that escaped gravity gliding and were thrust westward with the emplacement of the Berkshire massif in Late Ordovician(?). (See Trips B-2 and B-6). Importantly, these rocks have facies characteristics of both the Dalton and Hoosac and therefore probably represent the westernmost facies in the Taconic depositional basin because the Hoosac-Dalton-Cheshire sequences are connected by sedimentary interfingering (Norton, 1969).

The original depositional basin of the Taconic allochthon rocks at this latitude, based on the admittedly insecure arguments above, should have been in excess of 70 km wide. Palinspastic reconstruction of the Berkshire massif (see Trip B-6) suggests that the Precambrian crystalline rocks of the Berkshire massif, in the Middle Ordovician, were very likely about 60 km wide and located about 21 km farther east than their present position with respect to the miogeocline. The entire Taconic sequence could not likely have been

deposited on the "basement" that was to become the Berkshire massif, as has generally been suggested (for example, Zen, 1967, 1968) because rocks of the Dalton-Cheshire-Stockbridge shelf sequence were deposited on at least the western 30 km of the gneiss (Trip B-6). Bird and Dewey (1970) suggested that much of the sequence was deposited to the east of the Grenville basement. The Taconic depositional basin probably was located largely to the east of the rocks making up the present Berkshire massif, and east of the Hoosac facies. This argument suggests that the root zone of the allochthon lies somewhere within the vicinity of the Hoosac-Rowe boundary east of the Berkshire massif (see Trip C-11). The Taconic rocks were probably deposited (initially) in an ensialic, evolving to ensimatic, basin, with graben and horst structure and basaltic volcanism (Bird and Dewey, 1970; Bird, 1975). Grenville gneissic components may have been derived largely from intrabasinal sources, as the spatial relationships of the Giddings Brook-Chatham and Rensselaer Plateau slices cited earlier require. If such a model is true and the comparison with Triassic rift basins is valid, the Rensselaer facies may have been deposited throughout a considerable period of time and may not be the oldest rocks of the allochthon as commonly assumed.

Metamorphic and tectonic events in the central Taconics of N.Y. and Mass.

Table 2 (reproduced from Ratcliffe and Harwood, 1975) presents the major tectonic features recognized in a 50 km east-west belt extending from Mt. Ida and the Giddings Brook slice eastward into the core of the Berkshire massif.

Structures associated with emplacement of the allochthon
D_1 - Phase A of Taconic orogeny

Large recumbent folds, such as Zen (1961) reported from the northern region of the allochthon, have not been found in the central Taconic region. However, Zen and Ratcliffe (1968), and Ratcliffe (1969, 1974a, 1974b) report the existence of prefoliation minor folds both in the autochthon and allochthon. Through recent mapping in the Chatham slice, Bahrami and Ratcliffe have noted that a wide range of bedding-cleavage intersections are found within individual outcrops. Steeply plunging, almost reclined axes of major and minor folds are characteristic of both autochthonous and allochthonous rocks. Figure 2 shows that a set of pre-foliation folds does exist in the Chatham slice, and quite probably this is the cause of the steep and erratic plunges noted. Rocks of the Giddings Brook slice (see Stop 1 and figs. 3 and 5) reveal similar steeply plunging F_2 fold structures. No evidence for truly recumbent folds has been found. Wildflysch-like conglomerates are found at the sole of the Giddings Brook (Stop 2) and Chatham slices (Stop 6).

Phase B of Taconic orogeny

Emplacement of the Everett slice (high Taconics) was marked by tectonic breccia zones that are distributed along the Everett-Walloomsac contact and locally between the Everett and Chatham slices (Stop 9). The emplacement of all of the Taconic slices at this latitude was premetamorphic, and no evidence

Deformational event	Number of fold system	Types of folds and areal extent	Important tectonic features	Metamorphic event	Important crystalloblastic and other structures	Igneous intrusion	Probable age of rocks in figure 1	Orogeny
D_6	F_6	North-south open folds of foliation locally recognized in Stockbridge valley	Northwest- and north-trending normal faults		Hematite-cemented breccias		Uncertain (Middle Devonian to Late Triassic)	
D_5	F_5	N 25°-40° E.-trending upright to northwest overturned folds of foliation, with axial planar slip or crenulation cleavage. Folds recognized throughout area of figure 1 west to Mount Ida in SW corner of Kinderhook 15-minute quadrangle N. Y., where Taconic unconformity is folded by N. 40° E. upright folds	Refolds thrust sheets and blastomylonitic foliation Chatham fault	M_2 Thermal maximum Acadian metamorphism	Crenulation of sillimanite alined in axial surface of F_4 folds, granulation of garnet and staurolite that includes F_4 foliation mylonite of Taconic foliation		Middle to Late Devonian (Ratcliffe 1969a, b, 1972) Ratcliffe, Bahrami (in press)	Acadian orogeny
D_4	F_4	Northwest-trending upright to southwest-overturned folds with axial planar slip, crenulation, and flow cleavage. Folds recognized throughout area of figure 1, west to Chatham, N. Y., in center of Kinderhook 15-minute quadrangle	Folds thrust sheets and blastomylonitic foliation resulting in local overturning of thrusts, northwest-trending high-angle reverse faults		Muscovite, biotite realined and recrystallized in axial surface foliation, coarse sillimanite crystallized in foliation. Garnet, staurolite include folded F_2 fabric, and blastomylonitic foliation		Middle to Late Devonian (Ratcliffe 1969a, b, 1972)	
			Granite crosscuts thrust fault and blastomylonitic foliation	Thermal maximum Taconic metamorphism	Granite lacks blastomylonitic foliation in country rocks	Granite stock, South Sandisfield quadrangle	Late Ordovician(?) (Harwood, 1972)	Phases of Taconic orogeny
D_3	F_3	Northwest-trending recumbent to strongly southwest-overturned folds of basement gneiss and large-scale southwestward thrusting of Precambrian rocks of Berkshire massif across autochthon. Fold and thrust style recognized from Windsor quadrangle, Massachusetts (Norton, 1969), south to Norfolk quadrangle, Connecticut (Harwood, unpub. data), along west front of Berkshire massif	Faulted recumbent folds and nappes, mylonite gneiss, blastomylonite associated with major thrusts Thrust sheets at June and Canaan Mountains transported with Berkshire massif		Alaskite has weakly developed blastomylonitic foliation but intrudes more highly cataclastic rock in fault zones, mylonite gneiss, blastomylonite has muscovite, biotite, hornblende with lepidoblastic texture, cataclasis of F_2 foliation, thrusting synmetamorphic	Alaskite sills in faults and magnetite mineralization	Synchronous with latest movements or thrusts (Late Ordovician?) Thrusting probably late Ordovician based on age of crosscutting granite	D
D_2	F_2	Isoclinal northeast-trending northwest-overturned to nearly recumbent folds with strong axial planar foliation which is dominant foliation in most autochthonous and allochthonous (Taconic) rocks, but not clearly present in Paleozoic rocks attached to Berkshire massif. Folds extend west to Mount Ida where unconformable beneath lowermost Devonian	Folding of Taconic thrust contacts, regional foliation and refolding of slump or soft-rock folds in Taconic allochthonous rocks	M_1	Lepidoblastic muscovite, chlorite, biotite and ilmenite in foliation, chloritoid, albite include foliation but are kinked by F_4 structures		Middle to Late Ordovician(?)	C Taconic orogeny
D_2(?)	F_2(?)	Folding and metamorphism of Lower Cambrian metasedimentary rocks attached to Berkshire massif and in independent thrust slices at June and Canaan Mountains	Coarse foliation or schistosity formed	M_1(?)	Muscovite, biotite lepidoblastic in schistosity		Time of metamorphism very uncertain depending upon original position of these rocks, and timing of tectonic events at that site (Middle Ordovician to Cambrian?)	(C)?
D_1	F_1	Intrafolial minor folds associated with Taconic thrust contacts. Soft rock or slump folds in Taconic allochthonous rocks, scale of pre-F_2 folds not determined but widespread, area shown in figure 1, west to Mount Ida	Emplacement of upper Taconic slices (here, Chatham and Everett slices) Emplacement of lower Taconic slices	No metamorphism recognized	Tectonic breccias with inclusions of Stockbridge Formation along thrusts (Zen and Ratcliffe, 1971) Wild-flysch-like sedimentary rocks along base of thrusts		Uncertain (Middle Ordovician?) Middle Ordovician (Zen, 1972b, table 1)	B A
		Middle Ordovician unconformity						
D_0		Warping of Lower Cambrian to Lower Ordovician carbonate shelf sequence, locally dips near vertical (Ratcliffe, 1969a); possible block faulting					Late Early to Middle Ordovician (Zen, 1972b, table 1)	Pre-Taconic disturbance
		Pre-Dalton unconformity						
$D_{p€}$	$F_{p€}$	Isoclinal east-west-trending folds with generally steeply dipping axial surfaces and strong axial planar foliation, deformation of all Precambrian rocks including granitic intrusions such as Tyringham Gneiss	Gneissosity in Precambrian rocks of Berkshire massif	$M_{p€}$	Diopside, sillimanite, hornblende, microcline, perthite formed in dynamothermal event	Granodiorite-quartz monzonite intrusions such as Tyringham Gneiss, syntectonic	Dynamothermal event and granite intrusion approximately 1.04 b.y. (Ratcliffe and Zartman, 1971)	Grenville orogeny
		Pre-Tyringham foliation						

Degree of basement participation and intensity of deformation increasing with time

Table 2. Chronology of tectonic events in Columbia County, New York, and adjacent Berkshire County, Massachusetts. Modified from Ratcliffe and Harwood (1975).

is known in support of Bird and Dewey's (1970) suggestion that the Rensselaer Plateau and higher slices might have been metamorphosed prior to emplacement.

Phase C of Taconic orogeny (D_2 and M_1 Taconic metamorphism)

Following emplacement of all slices, regional dynamothermal metamorphism occurred, and a slaty cleavage or true axial planar foliation (S_2) formed in the rocks from the vicinity of Mt. Ida eastward into the area of the Berkshire massif and presumably beyond. In the low grade rocks, fine-grained sericite, chlorite, and lenticular quartz define the slaty cleavage. Small, round blebs of chlorite with 001 cleavage subparallel to bedding are ubiquitous in the low-grade rock and may be retrograded detrital biotite or diagenetic chlorite. However, lepidoblastic grains are not developed parallel to beds. Sandstone and siltstone dikes have not been found parallel to S_2, and no evidence thus far indicates that tectonic dewatering was an important mechanism in the formation of the Taconic slaty cleavage. Large finite strain is indicated by flattened pebbles that lie within the slaty cleavage. Locally, intense transposition structures are developed, and false bedding is common, particularly in laminated slates and some quartzites. Taconic thrust contacts of the Giddings Brook slice (Zen, 1961; Potter, 1972), Chatham (Ratcliffe, 1974a; Ratcliffe and Bahrami, in press), and Everett slices (Zen and Ratcliffe, 1968; Ratcliffe, 1968, 1974a, 1974b) were cross-foliated and folded during the D_2-M_1 metamorphic event to produce F_2 Taconic folds on a regional scale.

Phase D of the Taconic orogeny

Emplacement of the slices of the Berkshire massif and large-scale, westward overthrusting was concommitant with metamorphism. Recumbent folds formed both in the autochthon and in gneissic rocks (see Trips B-2 and B-6 for further amplification).

Acadian orogeny

Post-Taconic foliation structures are common throughout this belt and increase both in intensity and degree of concommitant mineral growth eastward. By using inclusion textures, we may delimit the approximate extent and character of the post-Taconic metamorphic imprint. East of the biotite isograd approximately at the New York State line post-S_2 mineral textures are abundant, indicating that the Acadian thermal overprint produced new mineral growth of muscovite (second generation with decussate texture), albite, chloritoid, biotite, garnet, and staurolite. It is fairly certain that the prominent mineral zonation is composite (polymetamorphic) and is dominantly controlled by the Acadian overprint in areas east of the biotite isograd. This probably explains the prevalence of Acadian K-Ar and Rb-Sr mineral ages (Zen, 1969) and the lack thus far of definitive Taconic mineral ages.

F_4 and F_5 folds are inconsistently developed and show contradicting relative ages from place to place. In eastern areas, the northeast-trending

refolds are the F_5 folds, whereas in the low Taconics east to the Stockbridge valley the northwest-trending refolds are the later folds.

The Chatham fault developed during the northeast-trending refolding episode, for it is refolded by northwest crenulation folds north of Chatham (Ratcliffe and Bahrami, in press). Locally, thrust faults with mylonitization of pre-existing foliation and chlorite-quartz-albite mineralization formed in sections of the Chatham slice containing massive quartzite and graywacke (Stops 7 and 8).

In the Hudson valley the Acadian deformation resulted in brittle fracture and development of crenulation cleavage and slip cleavage in Taconic rocks, and flexural slip folds of the Devonian rocks with numerous bedding-plane and low-angle detachment thrusts. Acadian structures become progressively metamorphic to the east, so that in the vicinity of the Berkshire massif and farther east, Barrovian-type, staurolite-kyanite-sillimanite metamorphism was characteristic of the Acadian dynamothermal event; rocks of similar grade could also have been formed during the Taconic orogeny, but confirmation of this point is thus far lacking in the Berkshires. If the model proposed for the Chatham thrust is correct (fig. 6), then Acadian deformation could have involved basement in the Hudson valley area.

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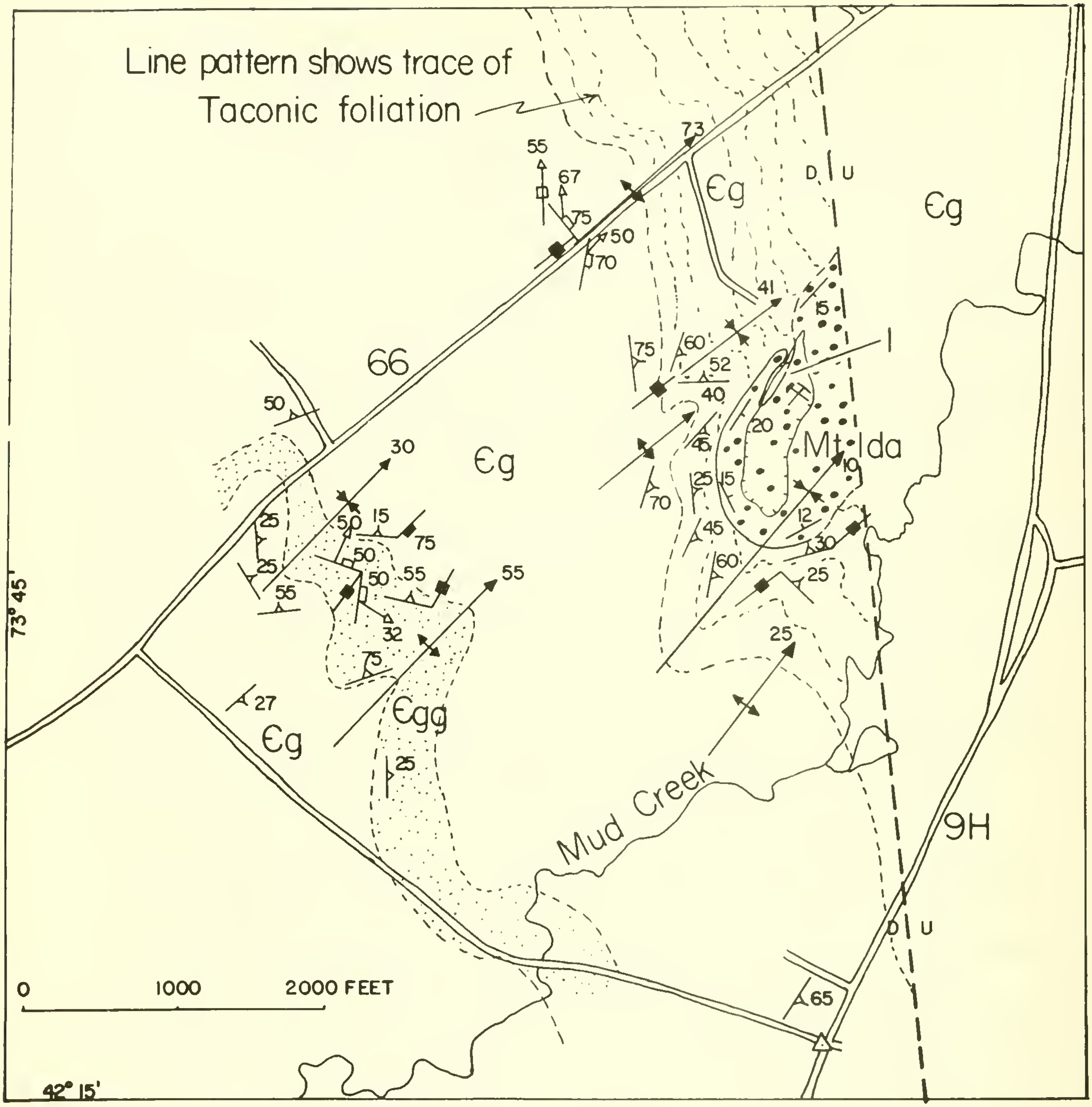
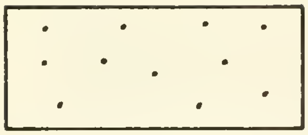


Figure 3. Geologic map of the Mount Ida area, Stop 1, showing areal distribution of Taconic foliation and attitude of post Lower Devonian (Acadian?) folds.

EXPLANATION

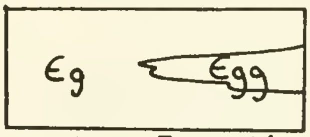
Silurian and Devonian



Dolostone and limestone

Taconic unconformity

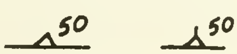
Lower Cambrian



Germantown Formation with gray-green graywacke Єgg and associated limestone conglomerate

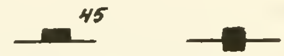
Contact between formations accurately located, approximately located

Strike and dip of bedding in Silurian and Devonian rocks



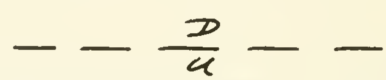
Strike and dip of pre-unconformity, penetrative foliation produced by parallel alignment of white mica, chlorite, and lenticular quartz, strike and dip of parallel bedding and foliation

Strike, dip of axial plane of pre-unconformity isoclinal folds having penetrative pre-unconformity foliation as axial plane. Arrow shows direction and amount of plunge of axis of fold (F_1 folds)



Strike and dip of inclined, and vertical post unconformity fracture cleavage that is axial plane to folds of foliation in pre-unconformity rocks, and approximately parallel to axial surface of folds in Silurian and Devonian rocks

Axial trace and approximate plunge of post unconformity folds of Taconic foliation (F_2 folds) and of initial folds of bedding post unconformity rocks



Approximate location of late high angle fault responsible for termination of Mt. Ida syncline, location and attitude conjectural

allochthon and surrounding autochthon in Bashbish Falls and Egremont quadrangles and adjacent areas, in Bird, J. M., ed., Guidebook for field trips in New York, Massachusetts, and Vermont: New England Intercollegiate Geol. Conf., 61st Ann. Mtg., Albany, N. Y., 1969, p. 3-1 to 3-41.

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Road Log

Assembly point Routes 66 and 9H, northeast of Hudson, New York (Brick Tavern), Stottville, N.Y., 7½ min. quadrangle.

Mileage

0.0 Head south on Rt. 66, 0.6 mile turn into Mt. Ida quarry at small yellow sign for Keil Contracting Corp. Drive to entrance to quarry and park.

Stop 1. Mt. Ida quarry and Taconic unconformity: discussion of Taconic and Acadian structures.

Mt. Ida is a small, fault-bounded syncline of uppermost Silurian and lowermost Devonian rocks, resting unconformably on slates of the Giddings Brook slice of the Taconic allochthon. Figure 3 shows the local geology.

Evidence for Taconic foliation and deformation

Dolostone of the Upper Silurian(?) Manlius limestone overlies slate of the Germantown "Formation" on the west wall of the quarry near the entrance. A limestone conglomerate 0-6 cm thick with chips of green slate forms the basal beds above the Taconic unconformity. The foliated chips contain a penetrative fabric outlined by oriented fine sericite, chlorite, and lenticular quartz that locally is normal to the postunconformity fracture cleavage, although most of the slate chips have been bodily rotated into parallelism with the Acadian cleavage. Although the pre-Manlius foliation commonly is parallel to the Acadian fracture cleavage, the foliated chips in the conglomerate document the existence

of a Taconic penetrative, low-grade metamorphic fabric. Figure 4 shows photomicrographs of slate chips and Acadian(?) cleavage in the basal conglomerate. Outside the quarry (fig. 3) the Taconic slaty cleavage has been mapped as broadly discordant with Acadian fracture cleavage and bedding in the Siluro-Devonian rocks.

Acadian deformation

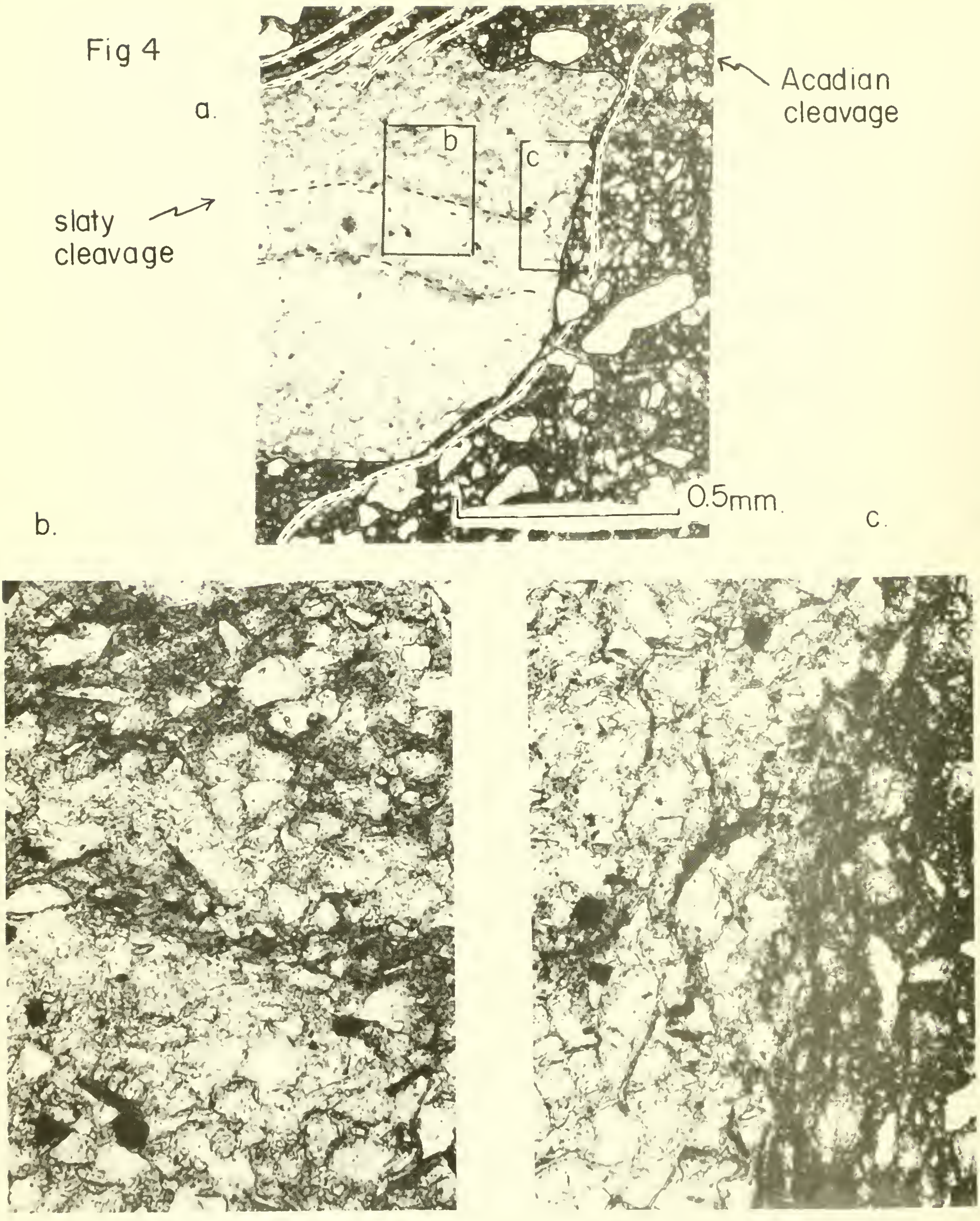
The Siluro-Devonian rocks and the unconformity are folded by broad flexural slip folds with abundant evidence of bedding plane slip. The age of this deformation is interpreted as Acadian because (1) folding in Devonian rocks dies out "upsection" in exposures west of the Hudson River, suggesting that the folding was pre-Upper Devonian, and (2) crenulation cleavage and post-Taconic foliation structures, when traced eastward, can be shown by metamorphic inclusion textures to be syn- or premetamorphic with respect to probable Acadian dynamothermal metamorphism (Ratcliffe, 1965).

A well developed fracture cleavage in the Siluro-Devonian rocks strikes N. 40° E. and dips steeply southeast. Immediately above the unconformity, thin seams of carbonate-rich silt intrude upward along the fracture cleavage for about a meter above as well as below the unconformity! Slate chips in the basal conglomerate also are bodily rotated without fracturing or bending, into a position subparallel to the Acadian fracture cleavage. Such rotation suggests a period of Acadian dewatering during or prior to flexural-slip folding. The origin of the alleged water (conate depositional water or sapprolitic from the slates) can also be debated on this outcrop.

Figure 4. Photomicrographs of foliated green slate chips in basal conglomerate of Late Silurian age, 6 inches above Taconic unconformity, Mt. Ida, Stop 1. Area of frames B and C outlined. Plane polarized light.

a. Shows slate chip surrounded by Upper Silurian quartz carbonate sand. Post-Lower Devonian fracture cleavage slopes from upper right to lower left and is outlined in white. Taconic slaty cleavage is horizontal.
b. Taconic slaty cleavage in slate chip, produced by fine grained sericite and chlorite concentrated in darker bands visible in (a) above.
c. Edge of slate chip, showing lepidoblastic chlorite and sericite and lenticular quartz re-oriented parallel to Acadian cleavage, shown in matrix. Texture implies that recrystallization of Taconic metamorphic minerals may have occurred locally in rocks as far west as the Hudson valley. Electron microprobe analysis of minerals will probably be necessary to demonstrate Acadian recrystallization rather than physical rotation of Taconic micas produced the new foliation. Micas similar to these are not found in the Silurian matrix.

Fig 4



Synopsis of Taconic and Acadian structures

Figure 5a, b, c, d shows structural data from the Mt. Ida area. In a, poles to foliation for the area of Figure 3 are shown, together with fold axes (open circles) of bedding with axial-planar slaty cleavage. The dashed small circle suggests the deformation path of these F_1 folds caused by Acadian rotation. Acadian fracture cleavage as measured in Devonian rocks (solid dots) and in Taconic slate (open dots) is shown in b. In c, the Taconic foliation measured beneath the unconformity has been restored by unfolding after taking out the plunge component of dip. The residual great circle configuration suggests the possibility of a weak, northwest-trending foldset (F_2? of diagram d). The synoptic diagram d shows approximate axial surfaces of the fold systems recognized. The pre-Manlius axial planes (slaty cleavage) were northeast-striking and steeply southeast-dipping, with fold axes (F_1) that plunge down the dip in nearly reclined folds prior to Late Silurian time. This pattern of steeply plunging F_1 folds is repeatedly seen throughout the Giddings Brook, Chatham, and Everett slices. Apparently these rocks contained rotated bedding prior to Taconic metamorphism and development of the slaty cleavage. This old structure may have formed during emplacement of the allochthon in the Middle Ordovician during a soft-rock (nonmetamorphic) event.

0.6 Log resumes at Rt. 66. Turn left (south) and proceed 3.4 miles south toward Hudson.

4.0 Turn right (west) on Rt. 23B at light and follow Rt. 23B through Hudson.

5.0 Turn left (south) on 9G and 23B. Follow signs to Rip Van Winkle bridge.

7.0 Roadcuts at Mount Merino 0.4 mile north of Rt. 23 intersection. Park off road or in parking lot east side of road near Rt. 23.

Stop 2. Wildflysch.

Zen (1961) proposed that the Giddings Brook and Sunset Lake slices of the Taconic allochthon were gravity slides and that the Forbes Hill Conglomerate of the northern Taconic region was a wildflysch facies that developed within the autochthon in response to the effects of the overriding submarine gravity slides (Zen, 1967). Subsequently, the wildflysch facies has been recognized along much of the western boundary of the Giddings Brook slice (Bird, 1969), and locally, along the eastern margin (Potter, 1972).

This stop is at one of the largest and best-exposed outcrops of the wildflysch in the Taconic region. It lies along the "front" of Mt. Merino, the "type-locality" of the Mt. Merino chert and shale (Ruedemann, 1942). Bird (1969) proposed that both Mt. Merino and a similar hill to the south called Mt. Tom (Mt. Thomas) are huge, detached blocks of the Giddings Brook slice within the wildflysch facies, because both hills are capped by Zone 12 and older rocks and, apparently, the wildflysch surrounds and projects under the hills. Bird also proposed that the Indian River - Mt. Merino facies is part of the Poultney Formation, not a member of the "Normanskill Formation", and that it is

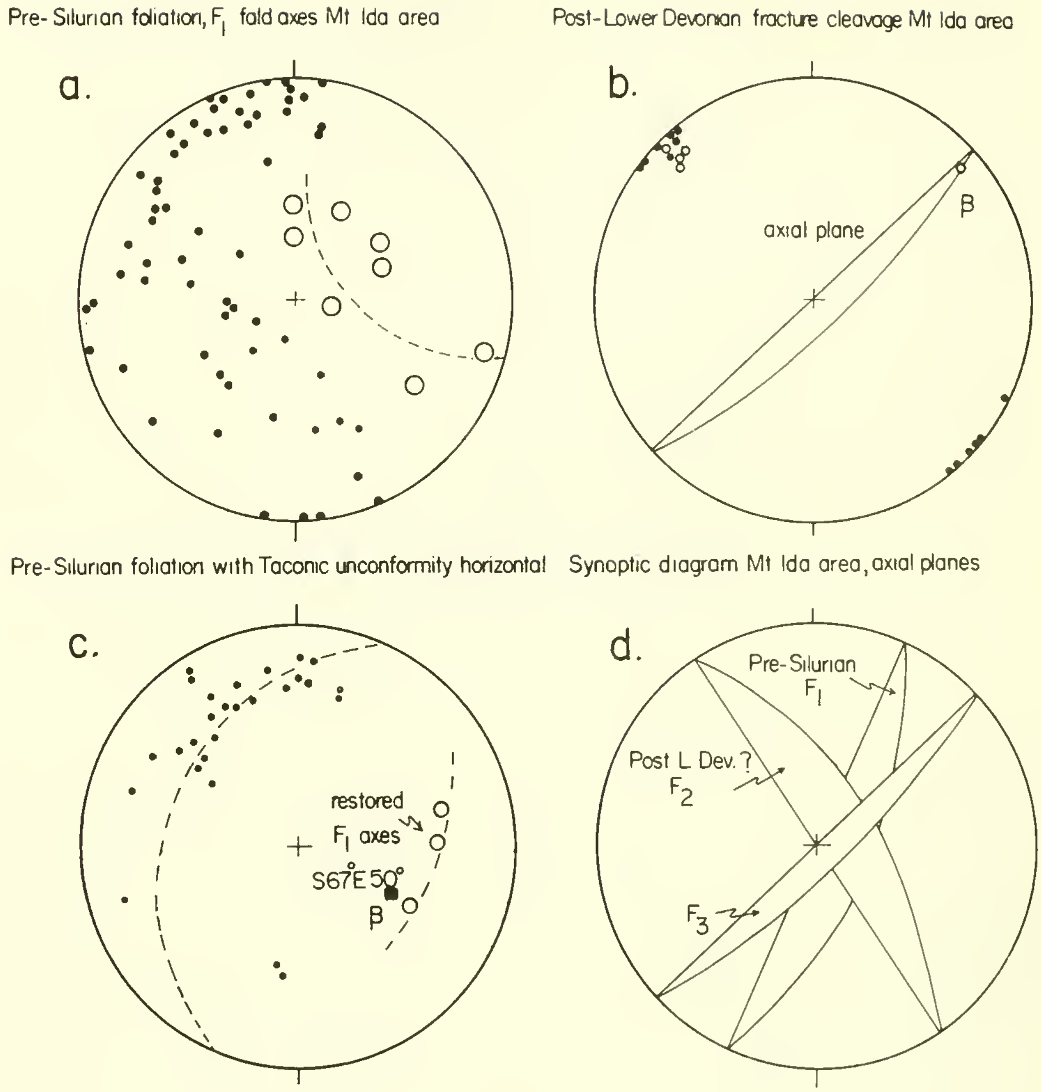


Figure 5. Lower hemisphere equal area projection of structural elements near Taconic unconformity, Mt. Ida, Stop 1.

a. Poles to pre-Silurian foliation area of Figure 3; large circles plunge of F_1 fold axes.

b. Acadian fracture cleavage in limestones open circles, crenulation or slip cleavage in Taconic slates solid dots, and axial surface and plunge of Acadian folds exposed in quarry.

c. Taconic foliation beneath 150 foot exposure of unconformity at west limb of syncline, with unconformity returned to horizontal, plunge component removed. Open circles F_1 fold axes restored, B pole to residual great circle, possible F_2 folds.

d. Synoptic diagram, axial surfaces Mt. Ida area. Nomenclature of F_1, F_2, F_3 is not intended to correlate with data on Table 1, where F_2 folds are equated with Taconic F_1 folds of this diagram.

entirely allochthonous, being the last facies to accumulate in the site of deposition of the Giddings Brook slice, before gravity sliding.

This exposure is within the belt of wildflysch that extends along the entire front of the Giddings Brook slice from southern Vermont to west of the Hudson Highlands (see Fisher, et al., 1971). In this region the wildflysch ranges from "soft-rock-deformed" Austin Glen graywacke and shale, to an extremely heterogeneous melange of the Austin Glen with included clasts of Giddings Brook slice lithologies. This exposure has inclusions of Mt. Merino chert, West Castleton - Hatch Hill(?) or Poutlney (Germantown - Stuyvesant of Fisher) bedded limestone and shale, and clasts of green Mettawee(?) shale. The rocks of this exposure are extremely complex and the trip leaders will attempt to point out relevant features. The exposure presents several questions: What are the cleavage relations of the shale matrix and clasts; did some of the clasts have a cleavage before incorporation into the melange: Note that some of the smaller Mt. Merino chert clasts must have been "soft" when deformed, and are not cleaved. Is all of the cleavage Taconic or is an Acadian cleavage also superimposed? Is the contact between the overlying, massive block of Mt. Merino chert and lighter colored melange, best seen on the southeastern end of the outcrop, characteristic of the overall nature of leading edge of the Giddings Brook slice?

The trip leaders believe that the overall character of the deformation of the wildflysch, the nature of the included clasts, and the distribution of the facies along the entire western edge of the Giddings Brook slice are ample evidence of Zen's original proposal. Detailed study of the wildflysch, particularly in the Hudson valley, has shown that apparently all the lithologies of the Giddings Brook slice, including Lower Cambrian fossil-bearing limestone clasts (Bird, 1963), and many lithologies of the carbonate sequence of the autochthon, especially Trenton-age facies such as the Rysedorph Hill Conglomerate, have been incorporated in the wildflysch which, everywhere, has a matrix of only the Austin Glen facies. No fossils other than Zone 13 graptolites have been found in the matrix. Therefore, emplacement of the Giddings Brook (and the very small Sunset Lake) slice as a huge, mostly submarine, gravity slide in Middle Trenton time is an established fact on the basis of the character and distribution of this most exotic, interesting facies.

7.2 Turn left onto Rt. 23 (east).

9.0 Junction Rt. 23 and 9 south. Continue on 23 and 9 east past south end of Becraft Mountain.

12.0 Road branches. Take left branch and follow Rt. 23 and sign for Taconic Parkway. Avoid Rt. 9 branch to right. Dangerous intersection.

12.2 Turn left (north) on Rt. 9H and 23 at light. Follow 9H and 23 north to Claverack.

16.0 Turn right (east) on Rt. 23 at light by Claverack Texaco station.

17.0 Turn left on Rt. 217, 1 mile east of Claverack and follow Rt. 217 3 miles north to Mellenville.

20.0 Mellenville, 500 ft. past Costa's store turn left up hill on unmarked road that leads to Columbia County Rt. 9 (do not cross creek into Philmont). Follow Col. Co. Rt. 9, 2 miles north to second crossroad north of Mellenville.

22.0 Turn right on paved crossroad, and follow road approximately 2 miles (go beneath RR overpass) to Stop 3. Park at top of hill.

Stop 3. Ghent Precambrian gneiss block and attached unconformable shelf sequence rocks, within the Chatham fault zone.

This remarkable and instructive exposure was recently discovered by Ratcliffe in 1973. Previous workers (Craddock, 1957; Fisher, et al., 1971 /after Craddock/) did not recognize the gneissic or shelf sequence rocks here but show the Chatham fault at this locality juxtaposing green slates and graywackes of the Nassau Formation (Fisher, et al., 1971).

Grenville-like alaskitic gneiss, hornblende granite gneiss, hornblende diopside plagioclase gneiss, amphibolite, diopside calc-silicate, all crossed by thin pegmatitic stringers, form an elongate sliver (1,000 feet long and 500 feet wide on the east side of the main Chatham fault). The western border of the gneiss (Chatham fault) is marked by mylonitic and highly cataclastic gneiss with east-dipping shear zones in the hanging wall and highly deformed, crenulated, and mylonitized olive-green slate and metaquartzite in the foot wall (seen at base of hill north of road).

Along its east and south border the gneiss is unconformably overlain by east-dipping, white to pinkish-gray vitreous quartzite (Poughquag?) with a basal quartz-pebble conglomerate south of the road and up the hill.

The quartzite, about 70 feet thick, is overlain by gray to white dolomite (Stissing Formation?) 100 feet thick, and this in turn is overlain by up to 600 feet of beige and orange-tan weathering dolostone with punky weathering quartzite typical of the Pine Plains Formation. If the lithologic correlations are correct, the Paleozoic shelf sequence of the block may range from Early to Late Cambrian in age. Along the east side of the sliver, gray to gray-green and olive slate of the Nassau Formation with graywacke beds up to 50 feet thick are thrust over and truncate the shelf rocks, and locally a small sliver of Walloomsac intervenes. Both rocks are cataclastic with gentle (35^{o}) east-dipping fractures and microfaults that cross the regional foliation.

Ratcliffe and Bahrami believe that the Chatham fault is a late tectonic feature based on the distinctive cataclasis of Taconic foliation in and near the fault zone along its entire width. They suggest that the Chatham fault is a major postmetamorphic thrust with a minimum of 3.5 km dip slip movement necessary to bring the gneissic basement

to its present position. The Ghent block may have been moved up in the fault zone by a stepwise set of faults as shown in Figure 6, or could have been plucked from a pre-Middle Ordovician age horst that was located somewhere east of the present trace of the fault. The Chatham fault actually is a very complex fault zone that contains large slivers of allochthonous Giddings Brook and Chatham slice rocks, all intermixed with blocks of Precambrian gneiss, Cambrian to Ordovician shelf sequence, and exogeosynclinal Middle Ordovician rocks from the autochthon, within a mylonitic matrix produced by granulation of all of the pre-existing metamorphic rocks.

A primary contact between the Giddings Brook slice and the Chatham slice has not been found and quite possibly does not exist. Postmetamorphic faults have not generally been recognized in the Taconics, but it appears that some of the major boundaries used to delineate separate slices of the Taconic allochthon (Zen, 1967) might be Acadian or younger structures. One might well wonder if this is Stissing Mountain in microcosm!

Continue east on dirt road.

25.0 Turn right at "T" intersection.

26.0 Park by bend in road one mile south at woods road leading to the west. (Outcrops are down the hill, 800 feet west of the road.)

Stop 4. Ashley Hill-like Limestone Conglomerate, West Castleton-Hatch Hill sequence, east of the Chatham fault.

Limestone as interformational and boulder conglomerate is widely developed west of the Chatham fault in rocks of the Giddings Brook slice (Zen, 1967; Bird and Rasetti, 1968) but only rarely east of the fault, as near Philmont (Weaver, 1957, p. 745) and this new locality discovered by Bahrami. As originally illustrated in Craddock (1957), the well known Ashley Hill Limestone Conglomerate (AHLC) (Dale, 1892) lies east of the Chatham fault. However, Zen (1967, p. 20) placed the Ashley Hill locality within the Giddings Brook slice (west of the Chatham fault) based on relocation of the fault by Talmadge (written communication to Zen, 1962).

This sliver, 4 km long and about 6 km wide, is located east of the west edge of the Chatham fault and illustrates imbrication into the Chatham fault of strata characteristic of the Giddings Brook slice.

The exposures here consist of limestone boulder conglomerate interbedded with black shale and dark-gray, fine-grained, well-bedded limestone, and punky-weathering, calcareous, crossbedded quartzite. Crossbeds indicate the east-dipping sequence is right-side-up. Green slates with quartzite and interbedded graywacke underlie the limestone and black slate. The contact or zone between these two lithologies is the "black-green boundary," a homotaxial surface that can be seen in this part of the section throughout the Giddings Brook slice (eg. Zen, 1964b). It is the defined contact between the base of the West Castleton-Hatch Hill sequence, which comprises the bulk of the fossiliferous

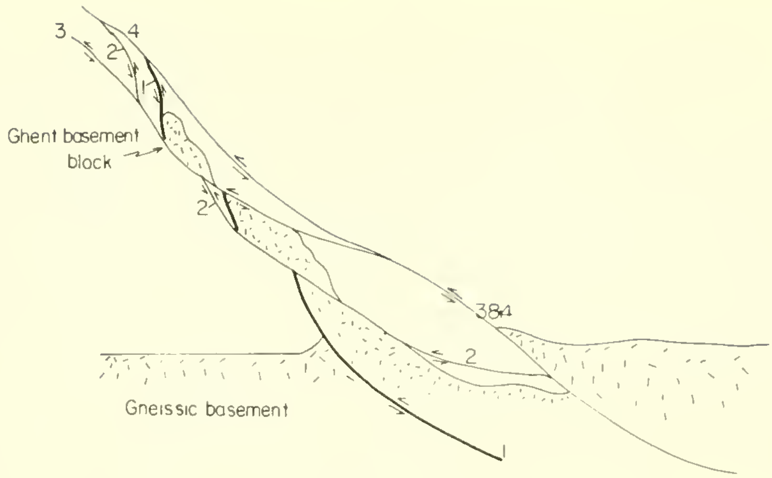


Figure 6. Diagram of possible mechanism for emplacement of the Ghent block. Numbers refer to sequentially developed faults. Slivers of shelf sequence rocks and Taconic allochthonous rocks would also move upward to produce a complex fault zone with interwoven slices of material initially plucked from widely different tectonic levels. Later faults need not offset basement but could be low angle detachment faults.

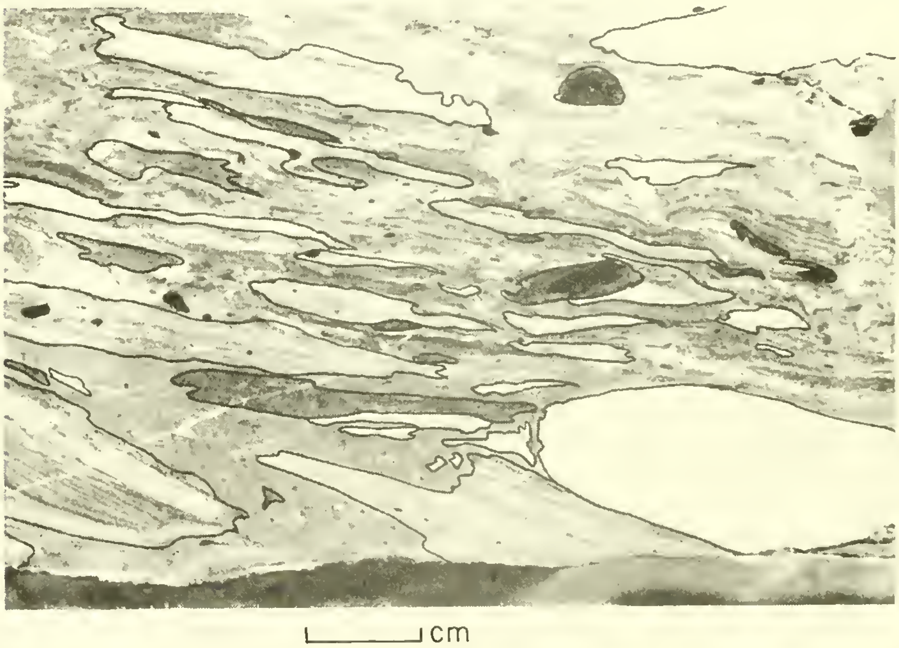


Figure 7. Sawn slab of wildflysch-like conglomerate from sole of Chatham slice, inclusions of gray-green slate, black slate in a nonbedded to well laminated gray slate matrix, Stop 6. Some of the more obvious inclusions are outlined in black. The localization of this feature at the Walloomsac-Nassau contact and the nature of the clasts suggest that the rocks of the Chatham slice were unlithified at the time of emplacement.

Cambrian strata, and the top of the Nassau or Bull Formation that comprises the bulk of the pre-fossiliferous, or Eocambrian strata. The green slate is Mettawee of the Nassau Formation; the black slate is of the West Castleton Formation. Lower Cambrian fossils occur in the Ashley Hill Limestone Conglomerate and the Mud Pond Quartzite (at Diamond Rock, Troy, N.Y.), which are facies that occur at, above, and below the "black-green boundary." These relations (and sources of data) are shown in Table 1, where a schematic columnar section that is "constructed" for Giddings Brook and Rensselaer Plateau slice rocks. It must be emphasized that, in contrast to the relatively straightforward stratigraphy upsection from the black-green boundary (Mt. Hamilton Group of Zen), we do not have a firm grasp of the actual stratigraphic relations downsection. For example, although Rensselaer Graywacke facies occurs in the lower part of the Nassau Formation in the type-locality of the Nassau quadrangle (Bird, 1961, 1962a, 1962b, 1969), the bulk of this facies, which is in the Rensselaer Plateau and Austerlitz masses, is within geometrically high structural units that have been telescoped from the original stratigraphic configurations and do not have the higher fossiliferous sections attached. Future work in the Taconics will certainly include determining the very complex palinspastic reconstructions of the Eocambrian-Cambrian rocks across such major tectonic features as the Chatham fault.

Continue southward on dirt road 0.4 mile.

26.4 Turn left (east) on Rt. 217.

27.0 Turn north on Taconic State Parkway (right turn under overpass).

28.0 Turn into rest area, east side of northbound lane of T.S.P. Stop 5 and lunch stop. Participants without lunch or in need of fuel may proceed northward on T.S.P. for 2 miles to Rigor Hill Rd. where there is a diner and Shell station. We will pick you up there in ½ hour.

Stop 5. Volcanic-polymict conglomerate and subgraywacke in the Nassau Formation. Park at turnoff from Taconic Parkway and walk 200 feet east. This is the first of three stops to examine Rensselaer-like facies of the Nassau in the Chatham slice.

Massive, dull-white weathering subgraywacke with a distinctive polymict basal conglomerate overlies gray, well-laminated slate. The pebbles in the conglomerate consist of (a) well rounded, garnet and zircon-bearing, white metaquartzite, (b) angular, dull, reddish-brown weathering, hematite-rich, glassy volcanic rock with irregular, globular textural variations suggestive of magma-filled vesicles, and microlites of plagioclase 0.1 mm in length that show a well defined flow structure, and (c) dull greenish-gray aphanitic or felsitic andesite(?) with microclites of plagioclase, quartz, and green hornblende(?). Rock (b) above probably is highly oxidized basaltic or andesitic scoria, perhaps the surface of andesitic flows (b) or of pyroclastic fragments.

The garnet-bearing metaquartzite is a typical Grenville rock in the Adirondacks, Green Mountains, and Berkshire massif (Washington Gneiss). A conglomerate similar to this, described by Balk (1953,

fig. 6), also contains black tuffaceous(?) fragments. Volcanic rocks are minor but important components of the Nassau sequence in the Rensselaer Plateau and Chatham slices and locally in the Giddings Brook slice (North Petersburg slice of Potter, 1972).

The mixture here of subgraywacke, well-rounded Grenville gneiss pebbles and fragmented, nonabraded volcanic fragments in a coarsely-graded lag deposit requires a special environment of sedimentation. Perhaps we are looking at sediments derived from fluvial environment, deposited in a basin made up of horsts and graben with active faults and basaltic volcanism. Bird (1975) suggested that the Rensselaer of the Nassau sequence represents a graben facies developed during the initial opening of the proto-Atlantic ocean basin (Bird and Dewey, 1970). In the context of that model Rensselaer Graywacke facies can be compared with those of the Triassic rocks of the Newark basin and Connecticut Valley. The Triassic rocks are fluviatile, shallow marine sediments and volcanic rocks, which developed synchronously with faulting attributable to the initial rifting of the present Atlantic ocean basin (Bird and Dewey, 1970).

30.0 Proceed north on T.S.P. for 2 miles to Rigor Hill Rd., where we will reassemble and pick up lunchless and/or gasless participants. Continue north on T.S.P. for 4.5 miles to intersection Rt. 203.

34.5 Exit for Rt. 203 east (turn right).

34.8 In 0.3 mile turn left at first intersection at Moorehouse Corner onto Columbia Co. Rt. 9 and follow paved road 2.5 miles to intersection with Columbia Co. Rt. 24.

37.3 Turn right on Col. Co. Rt. 24 at "T" intersection. Follow Rt. 24 0.9 mile east. Park before first barn on the right.

38.2 Stop 6. Nassau-Walloomsac "Taconic thrust" contact and wildflysch-like micromélange at sole of Chatham fault, Sheep Hole, Indian Brook, Red Rock.

Exposures of black Walloomsac in the brook are well-foliated with folds of bedding and bedding cleavage lineations plunging southeast at about 25°. Upstream, green phyllite of the Nassau overlies the black slate. Near the contact a zone several meters wide of mixed rock intervenes. Fragments of black slate in peculiar open ended forms float in a matrix of green slate, and both host and inclusions are strongly crossfoliated (fig. 7).

Unlike conventional wildflysch, this mélange consists of black chips (probably Trenton) in a green matrix probably of earliest early Cambrian or Eocambrian age. Locally this texture may be reversed with black and green chips set in a predominantly black Walloomsac matrix. This curious deposit has only been found at three localities, all within several meters of the sole of the Chatham slice. Although this is not a normal sedimentary-tectonic deposit, it could represent a zone of disarticulation and comingling of rocks within a zone of intense differential flow, formed at the sole of the Chatham thrust coeval with wildflysch produced

to the west at the leading edge of the slide. The encorporation of fragments of both rocks in the mélange without a cataclastic (brittle) fabric suggests soft rock deformation. This is the best evidence known to the authors suggesting that the Chatham slice was emplaced by a soft rock gravity slide mechanism similar to the Giddings Brook slice.

The contrast between the Chatham thrust (Stop 3) and this Taconic thrust contact is striking. The bulk of the contacts of the Chatham and Everett slices with the autochthon are premetamorphic.

Continue 0.6 mile east on Col. Co. Rt. 24 to Red Rock.

36.8 Turn right onto Macedonia Rd. Follow 0.8 mile to first major stream crossing.

39.6 Stop 7. Rensselaer graywacke contact with purple slate of the Nassau.

Large exposures of typical Rensselaer graywacke in the streambed overlie purple slate at the base of the waterfalls. The contact locally is a fault that is marked by pods of bull quartz, with alkali feldspar and veinlets of chlorite. These postmetamorphic faults are very common wherever massive units such as graywacke and/or quartzite are in contact with more ductile units. Two hundred feet north of the falls the purple slate is in sedimentary contact with the graywacke. Although these mylonitic, mineralized zones are spectacular and suggest mega-tectonic activity, usually they involve insignificant throw. This kind of deformation and mylonitization on low angle thrusts probably is the result of "bedding plane slip" during Acadian folding of rock sequences having high ductility contracts. In thin section irregular veinlets of chlorite, often wedge shaped, crosscut the older foliation. The chlorite commonly has a strong crossfiber fabric not related to either the older or later foliations. This behavior contrasts markedly with the similar style of folding and penetrative deformation active during Taconic metamorphism, as seen at the next stop.

39.7 Continue south on Macedonia Rd. 0.1 mile. Bear right staying on Macedonia Rd.

39.8 Intersection of Stonewall Rd. Continue on Macedonia Rd. for 1 mile.

40.8 Turn left at "T" intersection onto Reed Rd.

42.5 Intersection with Rt. 203. Turn left onto dirt road (Big Wood Rd.) that parallels 203.

42.9 Park before town dump.

Stop 8. Graywacke conglomerate and postfoliation thrust fault, west edge of the Austerlitz outlier of the Rensselaer graywacke.

Park on old Rt. 203 near gravel pit and walk up slope to north, 2100 feet to outcrop of graywacke at west draining brook. From this point head N. 70° E up slopes toward north end of hill 1356.

Large cliff exposure (Balk, 1953, Fig. 5, p. 825) exposes a 25 foot thick lens of the coarse graywacke conglomerate overlying a dark-gray to purplish-gray slate on an east dipping thrust fault. The boulders are flattened in the plane of the slaty cleavage (Taconic foliation). Numerous nearly reclined "S" folds of foliation and secondary segregations of quartz and veinlets of dark-green chlorite, quartz, and pink alkali feldspar form a mylonitic zone, with a downdip mineral lineation. Postfoliation thrust faults and mylonite zones with strong downdip chlorite streaking and bull quartz resemble some of the exposures cited by Potter (1972) as evidence for mylonitization along the sole of the north Petersburg and Rensselaer Plateau thrusts.

The conglomerate is polymict with Grenville(?) gneissic boulders, white calcite marble, layered amphibolite, black chert, reddish quartzite, and numerous large chocolate-brown weathering calcarbonate-rich rock. Fragments of Rensselaer graywacke and one small cobble of graywacke conglomerate can be found. Locally, diabasic volcanic fragments (Balk, 1953, Pl. b, fig. 3) identical to the basalt flows near Fog Hill in the State Line quadrangle (Ratcliffe, 1968, 1974a) are included in the graywacke, indicating the contemparaneity of the volcanism with deposition.

The brown weathering limey cobbles are not represented anywhere in the sub-Rensselaer stratigraphy of either the Giddings Brook or Chatham slices or inthe shelf sequence. They could represent a fine grained retrograded Precambrian marble or a limestone of unknown origin.

The mixture of autoclastic debris (graywacke clasts, limestone?, and volcanics) with probable Grenville clasts suggests that the Rensselaer may indeed have been deposited as a "Newark like" graben facies attendant to late Precambrian rifting as Bird (1975) suggested. The extreme coarseness of this deposit and the heterogeneous mixture suggest some kind of debris flow mechanism.

Fine grained sedimentary dikes (of graywacke) irregularly intrude gray-green slates in excellent exposures 300 feet along the slope to the north. This probably resulted from abnormal pore pressures resulting from rapid sediment loading of coarse debris flows onto uncompacted shales and graywackes. Balk also reported sedimentary dikes in the Rensselaer northeast of Troy (Balk, 1953, pl. 3).

43.3 Continue east and turn left on Rt. 203 in 0.4 mile.

45.9 Intersection with Rt. 22 at Austerlitz. Turn right (south). Hills east of Rt. 22 are sapped by rocks of the Everett slice that overlie the Walloomsac. West of the highway fault slices of Everett cap the higher hills, and locally the Everett extends down to the road. The belt of Walloomsac in the valley is parautochthonous and has been thrust westward over the Chatham slice.

47.0 A marked bench 3/4 way up the slopes to the east of the road marks the location of a prominent carbonate tectonic breccia zone at the sole of the Everett slice. A detailed map of this locality was published in Zen and Ratcliffe (1966).

48.9 Intersection with Rt. 71. Bear left on 71 toward Great Barrington. Hills to the right are imbricate and folded slices of Everett interleaved with tectonic breccia and slivers of parautochthonous Walloomsac.

50.3 Turn right onto Overlook Rd., opposite the Green River Inn.

51.9 Stop 9. Tectonic breccia at sole of Everett slice. Discussion of the Everett slice. Park near T.V. tower. Exposures are in woods above pasture east of road.

Tectonic breccias at the base of the Everett slice include fragments of Stockbridge carbonate rocks, Walloomsac limestone and phyllite, and fragments of Everett intermixed in a complex zone locally up to 100 feet thick. Zen and Ratcliffe (1966) believe these inclusions were plucked from the underlying autochthon during emplacement of the Everett slice and physically dragged along beneath advancing thrust. All units of the Stockbridge and Walloomsac are represented in these breccia zones, although no autochthonous rocks older than unit a of the Stockbridge have been found. This is the best exposed large outcropping of the breccia in southwestern Massachusetts. Zen's 1969 N.E.I.G.C. trip visited this locality (Stop A6). Detailed geology is available in Zen and Ratcliffe (1971). Zen has new information about the mineral assemblages and their isotopically determined ages that he has kindly consented to discuss at this stop.

The breccia zone is exposed east of the road in the core of and on the eastern limb of a northeast-plunging anticline (fig. 2). West of the road green Everett phyllite is found in the core of the complementary syncline rimmed by the tectonic breccia. This western belt of breccia traces northward into the State Line quadrangle beneath a belt of Everett that is the highest of a series of imbricate slices at the leading edge of the Everett slice. The anticline-syncline pair have an axial surface that strikes N. 10° E. and dips 60° to 65° southeast parallel to regional foliation. Bedding cleavage lineations plunge downdip or to the northeast in a style similar to the F_1 folds seen to the west.

Examination of the contacts of the marble inclusions with the Everett and the Walloomsac shows that the earliest foliation parallels the contacts on the limbs of minor folds and crosscuts the marble-phyllite contact on the noses of folds. The fabric here and elsewhere in the Everett-related breccia zones differs from that found in the Chatham fault, Stop 3, where contacts of limestone and other inclusions postdate the foliation.

Secondary plications of foliation and a fairly well developed second foliation, dipping more steeply east than the first foliation, is not developed or concentrated at the contact of the carbonate inclusions. Ratcliffe's interpretation, which may differ from that of Zen (1969), is that the breccia zone was emplaced prior to rather than synchronous with the structural event that produced the first regional foliation (Taconic). Ratcliffe believes the overall highly folded geometry of the breccia zone about the early axial planes requires this interpretation.

Metamorphic textures in the Everett at this locality (Zen, 1969) and on strike in the State Line quadrangle (Ratcliffe, 1965, 1969) indicate that second generation muscovite, zoned albite, chloritoid, and zoned garnets poikiloblastically include both an older foliation and micro-crenulations of it, and are locally crystallized across micro-offsets of the slip cleavage. East of the garnet isograd, new biotite and muscovite lie in new foliations that are the axial plane of the F_4 and F_5 folds of Table 2.

Ratcliffe believes that these textures indicate a petrographically definable zone of Acadian overprinting. The intensity of the over-printing, as indicated by the mineral growth textures, suggests that the K-Ar and Rb-Sr ages available probably record the effect of the Acadian thermal history, and older Taconic minerals, if preserved at all, should yield hybrid ages.

Excellent exposures illustrating the polymict nature of the tectonic breccia may be seen east of the pasture in the west-facing ledges. Magnificent 3-dimensional, crawl-around exposures of marble inclusions are preserved in small caves near the south end of the ledges.

End of trip.

Return to Rt. 71. Turn right and follow to Rt. 23; turn left on Rt. 23. Turn left on Rt. 9 for Great Barrington.

Trip B1, C1

SELECTED LOCALITIES IN THE TACONICS AND THEIR IMPLICATIONS FOR THE PLATE TECTONIC ORIGIN OF THE TACONIC REGION

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Introduction

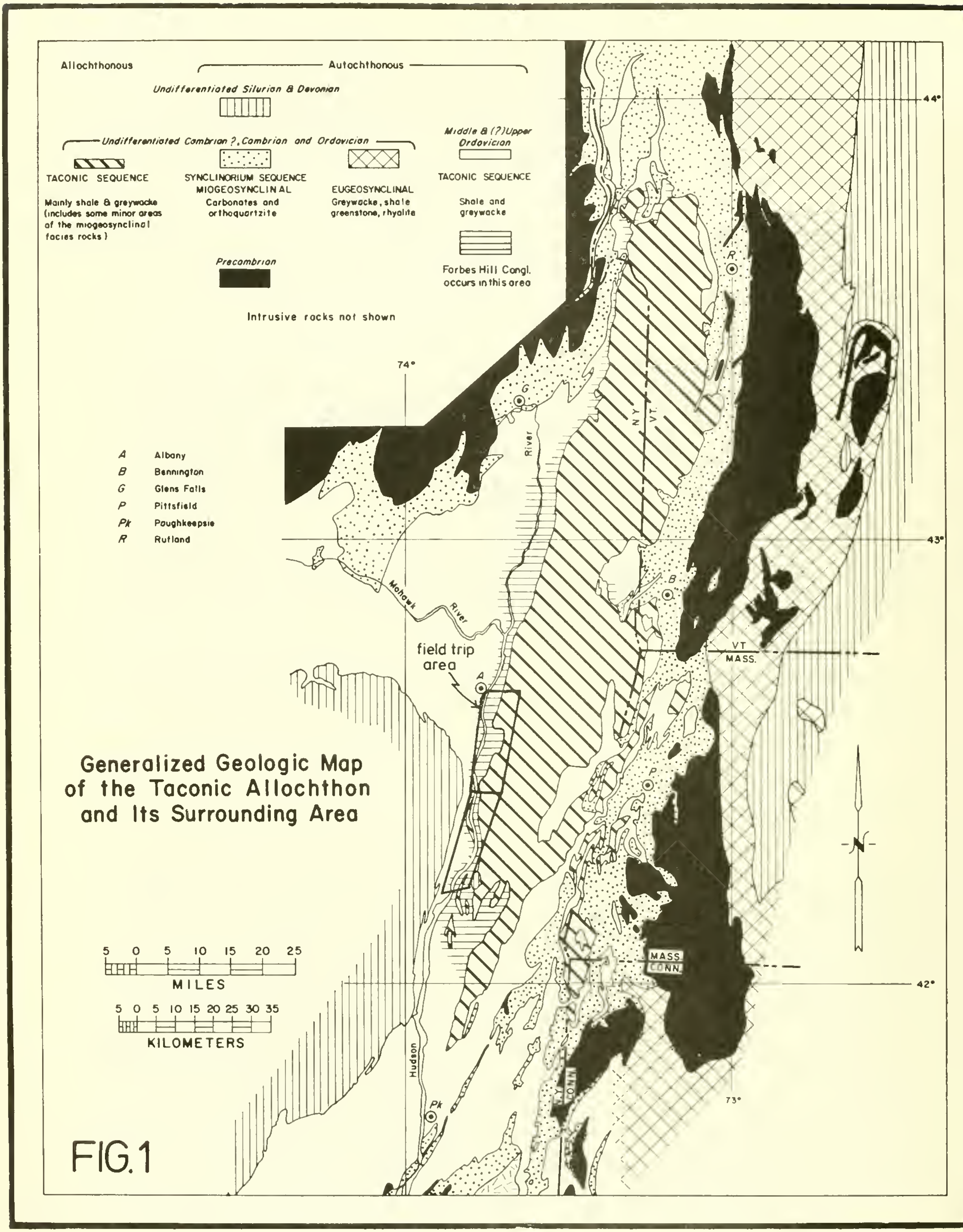
With the advent of plate tectonics concepts, orogenic belts have become more understandable in actualistic terms. Present plate margins provide examples of styles and histories of deformational belts which extend back in time to the early Mesozoic, as a consequence of the evolution of the present plate population. The magnetic anomaly record of the sea-floors provides a kinematic basis of plate reconstructions.

Because the present plate population has evolved since early Mesozoic time, there is no direct way of evaluating Paleozoic and older orogenic belts in the context of evolving plates and plate margins. Therefore, at best, we can make only second-order deductions about pre-Mesozoic orogenic mechanisms in the context of plate evolution.

We (Bird and Dewey, 1970) have attempted to do this by integrating the geologic history of the Taconic region into a plate evolution model for the northern Appalachians. The purpose of this trip is to examine a few of the rocks of the Taconic region that provide some basis for that model.

Geologic History of the Taconic Region

The Taconic region (Figure 1) is bracketed by the Precambrian-age Grenville basement igneous-metamorphic complex of the Adirondacks to the northwest, and by the Berkshire/Green Mountain massif to the east. The basement is overlain by a Lower Paleozoic platform assemblage having a Lower Cambrian to Lower Ordovician basal transgression (Cheshire-Potsdam) capped by a Lower Cambrian to Middle Ordovician carbonate platform sequence. This is followed by an autochthonous Middle to Late Ordovician exogeosynclinal flysch basin, into which the various structural members of the Taconic allochthon were emplaced. This assemblage of autochthonous and allochthonous rocks is overstepped by a progressively eastward-developed unconformity between Ordovician and Silurian strata. The Silurian/Devonian sequence of the Heldebergs records Acadian deformation, superimposed on the entire region. The carbonates of the Heldeberg sequence pass upward into the Middle to Late Devonian clastic assemblage of the Catskill region, a typical molasse facies. The Taconic region is, essentially, a synclinal belt between the Hudson/Champlain Valley and the Berkshire/Green Mountains and can be divided into four major structural/stratigraphic features:



Allochthonous
Autochthonous
Undifferentiated Silurian & Devonian
Undifferentiated Cambrian ?, Cambrian and Ordovician
Middle & (?)Upper Ordovician
TACONIC SEQUENCE
Mainly shale & greywacke (includes some minor areas of the miogeosynclinal facies rocks)
SYNCLINORIUM SEQUENCE MIOGEOSYNCLINAL
Carbonates and orthoquartzite
EUGEOSYNCLINAL
Greywacke, shale greenstone, rhyolite
TACONIC SEQUENCE
Shale and greywacke
Forbes Hill Congl. occurs in this area
Precambrian
Intrusive rocks not shown
A Albany
B Bennington
G Glens Falls
P Pittsfield
Pk Poughkeepsie
R Rutland
Generalized Geologic Map of the Taconic Allochthon and Its Surrounding Area
field trip area
Mohawk River
River
Hudson
NY VT
VT MASS.
MASS. CONN.
NY CONN.
74°
73°
44°
43°
42°
5 0 5 10 15 20 25
MILES
5 0 5 10 15 20 25 30 35
KILOMETERS
N

FIG. 1

1) The miogeoclinal basal transgression - carbonate platform wedge, 2) The Trenton-age Balmville Limestone (unconformable on the platform via the sub-Trenton/Black River unconformity) and Normanskill shale/Austin Glen flysch facies which changes into the Trenton limestone facies in the Mohawk Valley region, 3) The Taconic allochthon, and 4) the mantling Silurian/Devonian carbonates exemplified by the relations at Mt. Ida (see trip A3) and Becraft Mountain. The boundaries and regional distribution of all these stratigraphic/structural assemblages were discussed in detail by Zen (1967) and Bird (1969).

Of these four assemblages, the Taconic allochthon is the most complex, and most instructive; it includes rocks transported from original sites to the east which have been obliterated by intense deformation and metamorphism.

The Taconic allochthon was emplaced in several stages, the first being the gravity sliding emplacement of the Giddings Brook-Sunset Lake slices (Zen, 1961, 1967; Bird, 1969) in post-Balmville, Zone 13 (Berry, 1960) time. (The Sunset Lake slice is very small; for the purpose of this article we include it with the term Giddings Brook slice). This event produced an exotic, wildflysch-like facies, by the overriding of the frontal erosional products of the advancing submarine slide. We will examine several exposures of this syntectonic facies, along the western "front" of the Giddings Brook slice; it provides the evidence for the time/space framework, and mechanical aspects of the gravity sliding. Zen (1967) provided a detailed, regional synthesis of the relations of this and the higher slices of the Taconic allochthon. The timing and structural "style" of emplacement of the higher slices remains as one of the key problems of Taconic geology (see trips B2, B4, B6 in this guidebook). The intensity and grade of Taconic metamorphism increases eastward, and strong Acadian deformation and biotite-grade metamorphism of the eastern Taconic region in southern Vermont increases to garnet and sillimanite-grade east of the Hudson Highlands in the southern Taconic region. The western limit of the Acadian deformation is along the eastern border of the Catskills. Because of the extent of this deformation and metamorphism, the most useful strata for reconstructing Taconic time/space relations are preserved in the northern regions of the autochthon and for the most part, in the Giddings Brook slice, because of their contained faunal assemblages and original depositional features. Therefore, for the purposes of this trip, and because of logistical constraints, we will concentrate on the western region of the Giddings Brook slice.

The Giddings Brook Slice

Following the recognition of gravity sliding of the initial Taconic allochthon (Zen, 1961), Rodgers (1968) showed that the eastward termination of the carbonate miogeocline, in the Milton region of Vermont, was a facies change into breccias, and then to shale (the Rugg Brook Conglomerate - St. Albans Slate). He pointed out that this facies change and the regional stratigraphy show that the eastern termination of the carbonates was a carbonate platform edge, with the deeper-water shale to the east being of an off-shelf rise environment.

Also it has been learned, principally through the detailed mapping of Zen (1961) and Theokritoff (1964), that the Giddings Brook stratigraphy extends from Eocambrian through to Middle Ordovician (Zone 12 of Berry) and that the time span of the fossiliferous portion matches that of the carbonate miogeocline (Zen, 1961; Bird and Rasetti, 1968; Bird, 1969). In the eastern region of the autochthon, the Lower Cambrian Cheshire Quartzite overlies the Dalton and Nickwaket; the Mudd Pond Quartzite of the Giddings Brook slice is the equivalent of the Cheshire (Bird and Rasetti, 1968). The Cheshire and Mudd Pond both contain the stratigraphically lowest Olenellus biozone fossils of the region, and mark the base of the fossiliferous strata of the autochthon and allochthon, respectively. The Dalton-Nickwaket facies and the Cavendish facies further east, resemble sub-Olenellus biozone facies of the Nassau (or Bull) Formation. The facies of the West Castleton - Hatch Hill and Poultney Formations of the Giddings Brook slice can be readily attributed to the starved rise region that must have existed off the carbonate bank. (We will examine several exposures of these facies). Essentially then, through the work of Zen, Rodgers, Bird, and Bird and Rasetti, the fossils, facies and regional stratigraphic and structural relations of the autochthon and gravity slide portion of the allochthon have been fitted back to pre-orogeny configurations of shelf and off-shelf assemblages which were to be telescoped in the early phases of the Taconic orogeny. These original configurations are illustrated in Figure 2.

The boundaries of the autochthon, the basal transgression on the basement, from Lower Cambrian on the east to the Upper Cambrian on the west at the Adirondacks, and the mantling Normanskill shale and graywacke, are well established (Figure 1). The rocks of the Giddings Brook slice, however, are completely detached from their original setting. Mapping in the Nassau Quadrangle shows that the Nassau Formation includes hundreds and perhaps thousands of feet of shale and siltstone beneath the lowest fossil-bearing strata (Mudd Pond Quartzite horizon). Lower down in the Nassau are graywacke and shale directly equivalent with the Rensselaer graywacke, and apparently, this was of a graben-horst non-marine environment (Bird and Dewey, 1970; Bird, 1975). We do not know what may have constituted the "base" of the Giddings Brook strata; the higher slices of the Taconic allochthon contain older sequences of strata that resemble or are equivalent to the lower parts of the Nassau (or Bull) Formation (see Zen, 1967, p. 15, Fig. 4). With the bulk of the Taconic allochthon in the higher slices, it is important to note that most of the allochthonous rocks are in fact pre-Olenellus in age and therefore properly assigned to the Precambrian. The fossiliferous strata, from near the top of the Nassau or Bull Formation up-section to the Indian River - Mt. Merino sequence, in the Giddings Brook, Bird Mountain (Zen, 1967) and perhaps in the Chatham slice (see Ratcliffe, trip A3, this guidebook), are probably no more than 1000 to 1500 feet thick (Fig. 2).

The "top" of the allochthonous sequence, the Indian River - Mt. Merino facies of the Poultney Formation (Bird, 1969), best seen in the Giddings Brook slice, is correlative with the sub-Trenton/Black River unconformity of the autochthon, and includes the last sediments deposited in the off-shelf region before gravity sliding. The Indian River and Mt. Merino are entirely allochthonous. The upper boundary of the Giddings Brook slice is a syn-gravity sliding facies - the Pawlet in the northern

FIG. 2 Eastern New York State

Schematic reconstruction of pre-"Taconian orogeny" stratigraphic relations

Thicknesses and distances variable or unkawn

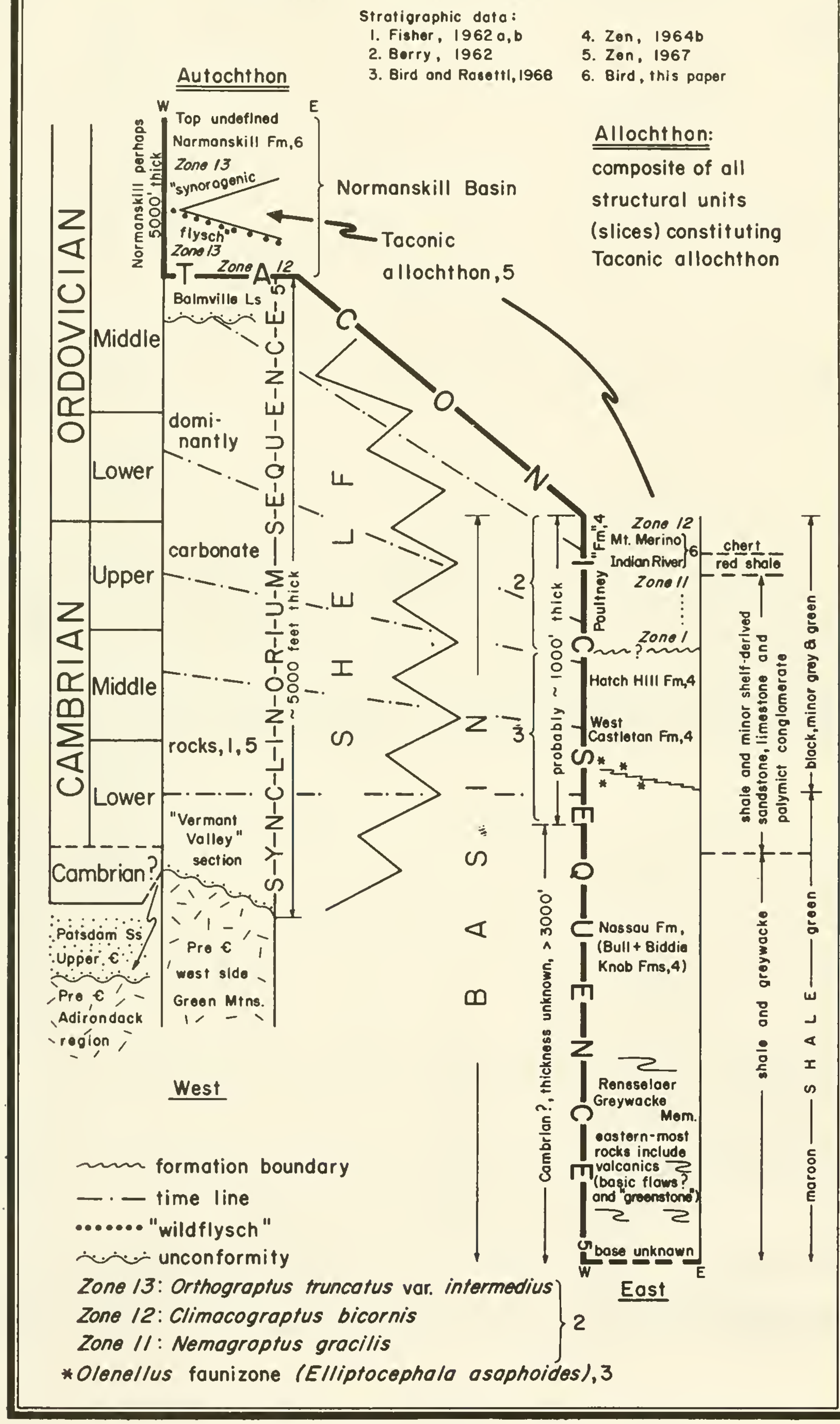
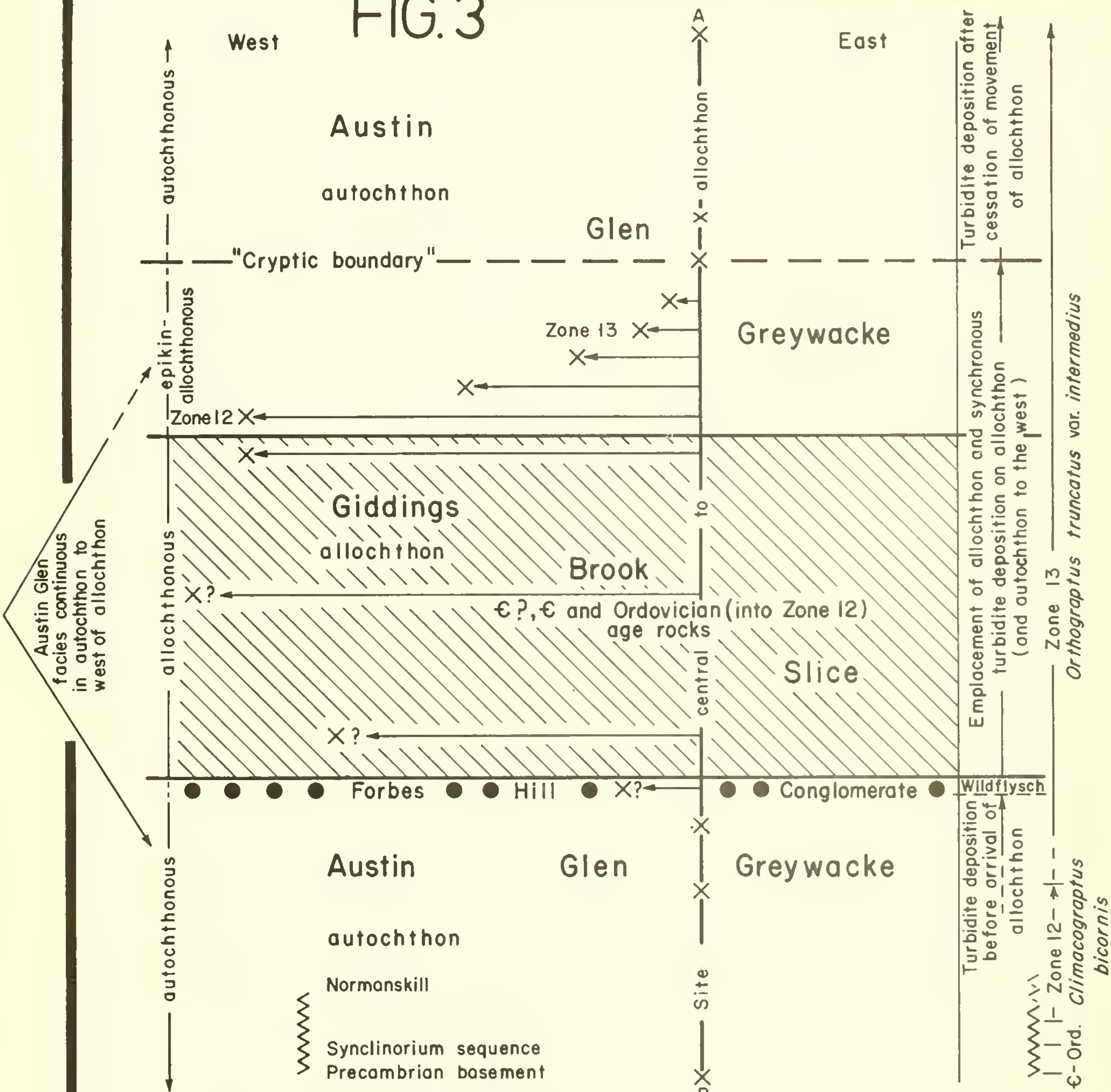


FIG. 3



Horizontal arrows indicate displacement of points X originally at vertical geographic reference line A-B, during emplacement of allochthon and synchronous sedimentation (Austin Glen Greywacke, Mem. 4 of Normanskill)

Hypothetical Section Illustrating Sedimentation on Moving Allochthon
Emplacement of Giddings Brook Slice into
Normanskill Sedimentary Basin

Scale: Schematic - thicknesses, and relative lateral positions and distances variable; horizontal boundaries assumed; Zig-Zag represents abbreviated section.

Taconics and the Austin Glen in the central and southern regions, which is either gradational with the upper Poultney, the Mt. Merino shale and chert, or has a deep-bite unconformity with rocks of the Giddings Brook slice as low as the Mettawee of the Nassau or Bull Formation (Zen, 1961; Theokritoff, 1964). These relations are described in detail by Bird (1969), and need not be recounted here except to point out that they indicate the gravity sliding was submarine for the most part and that the Pawlet /Austin Glen was deposited, in some places, upon the moving allochthon as well as in the basin into which the gravity slide was moving. These relations are illustrated in Figure 3. Deposition of this flysch facies continued on after cessation of gravity sliding, as indicated by distribution of the flysch to the west of the allochthon and above the wildflysch. To the southwest, at Illinois Mountain, a coarse molasse facies (the Illinois Mountain quartzite of Bird and Dewey, 1970) bears upper Ordovician fossils (D.W. Fisher, pers. comm., 1969) and almost certainly represents a syntectonic facies developed during the later Ordovician thrusting of the higher slices of the "high" Taconics (Fig. 4C).

Model for Evolution of the Taconic Region

We have proposed that the Taconic region was a segment of the continental margin of an evolving Paleozoic ocean system (Bird and Dewey, 1970). The pre-deformation stratigraphic framework represents the structural and depositional evolution, from late Precambrian to Middle Ordovician, of the continental margin of the opening proto-Atlantic, or Appalachian ocean. The emplacement of the Taconic allochthon was an effect of the conversion of that margin to an Andean-like orogenic system. The structural evolution of the region culminated in the Acadian, with suturing by continent to continent collision, during the final stages of closing of the ocean. Figure 4 illustrates, schematically, a model based on lithosphere plate evolution.

On the eastern side of the Taconic region, and on the east limb of the Green Mountain anticlinorium are sedimentary sequences that underlie the Cheshire quartzite, the Lower Cambrian fossil-bearing basal transgression facies of the carbonate miogeocline. These include the Dalton (Mendon) and Nickwaket. Within the Giddings Brook slice, far beneath the lowest fossil-bearing unit (the Mudd Pond Quartzite zone of the Bull or Nassau Formation, and the equivalent of the Cheshire Quartzite of the autochthon) are Rensselaer graywacke-like strata, comparable with the bulk of the facies in the Rensselaer slice. When reconstructed to a pre-orogenic configuration (Figure 4A), these early facies occupy a position consistent with a model of Late Precambrian rifting of a continental mass. Basic volcanics occur in the Rensselaer facies and also in the Cavendish (the Tibbett Hill Member) and Pinnacle of the East Vermont sequence and we suggested (Bird and Dewey, 1970) that this assemblage accumulated in extensional graben structures during the earliest phases of continental rifting (see stop 8, trip A3, this guidebook). The Appalachian ocean shelf or miogeocline and continental rise (original position of Bull - West Castleton - Hatch Hill - Poultney sequence) evolved synchronously with the late Precambrian - Early Paleozoic opening of the Appalachian Atlantic (or proto-Atlantic) Ocean. Grenville basement rocks must have occurred under this segment of rise sediments, as indicated by

the existence of this basement to the east, in the Chester Dome. However, in the context of continental lithosphere rifting it is likely that a full thickness of continental basement did not exist oceanward, beyond the shelf edge. This boundary, persistent through the time of evolution of the carbonate platform and rise, later became a significant tectonic and stratigraphic break that influenced the development of the major, regional structural features of the Taconic allochthon.

In addition to their synchroneity with the carbonate platform sediments, the West Castleton through Poultney sediments contain varieties of facies and sedimentary structures that are evidence of a starved rise region of deposition (Bird and Rasetti, 1968). We will examine some of these facies and structures at Schodack Landing and Judson Point. Also, the absence of significant amounts of coarse terriginous sediments in this sequence indicates not only the extent of the carbonate platform and effective terriginous sediment-blocking, but also that no significant relief developed in the interior craton during the time of deposition. It is very important to recognize that the coarse clastics of the pre-Olenellus Taconic sequence must have been underneath and perhaps further oceanward from the rise assemblage, as indicated in Figure 4A. On the other hand, very coarse and angular limestone breccias and conglomerates occur in the Upper Nassau and West Castleton. These chaotic layers, usually having a rounded-sand and carbonate-cement matrix, are interbedded with black and green shale, and clearly, are exotic slump deposits from a shallow carbonate-bank edge environment. Also, further up-section, in the Hatch Hill, ferruginous quartzite beds in shale, up to several meters thick, are best attributed to sand-flow and turbidite mechanisms, again from a shallow shelf environment. There is also a wealth of turbiditic, sandy, silty and carbonate beds and minor conglomerates in the Poultney, particularly in the Deepkill and Schaghticoke facies, that are evidence of continued starved, off-shelf depositional environments, contemporaneous with the Early and Middle Ordovician platform sediments.

During early Middle Ordovician time, the miogeocline-rise couple was affected by uplift and block faulting. An extensive erosional surface developed which was to become the sub-Trenton/Black River unconformity (see Zen, 1967, for a detailed account of this phase of Taconic evolution). This event was coeval with development of the Indian River facies of the Poultney; the red shales of the Indian River are thought to be derived from a terra rosa soil horizon on the erosional surface (Bird and Dewey, 1970). The Indian River red shales pass upward into the green and black shales and cherts of the Mt. Merino. Detailed petrographic study (Bird and Lang, unpublished) indicates clearly a volcanogenic origin for much of the Indian River - Mt. Merino chert. These chert facies are correlative with the Ammonoosuc volcanics, of Vermont and Massachusetts, and pass upward, locally, into shale and graywacke beds of the Pawlet.

As shown in Figures 2, 3, and 4B, the emplacement of the Giddings Brook slice was contemporaneous with the westward encroachment of flysch into the Mt. Merino environment, the sagging of the miogeocline and attendant westward migration of the Balmville carbonate facies to become the Trenton limestone of the Mohawk Valley, and deposition of Normanskill shale. The Pawlet - Austin Glen Graywacke facies overwhelmed the subsiding continental margin, contemporaneously with a major relief inversion. The

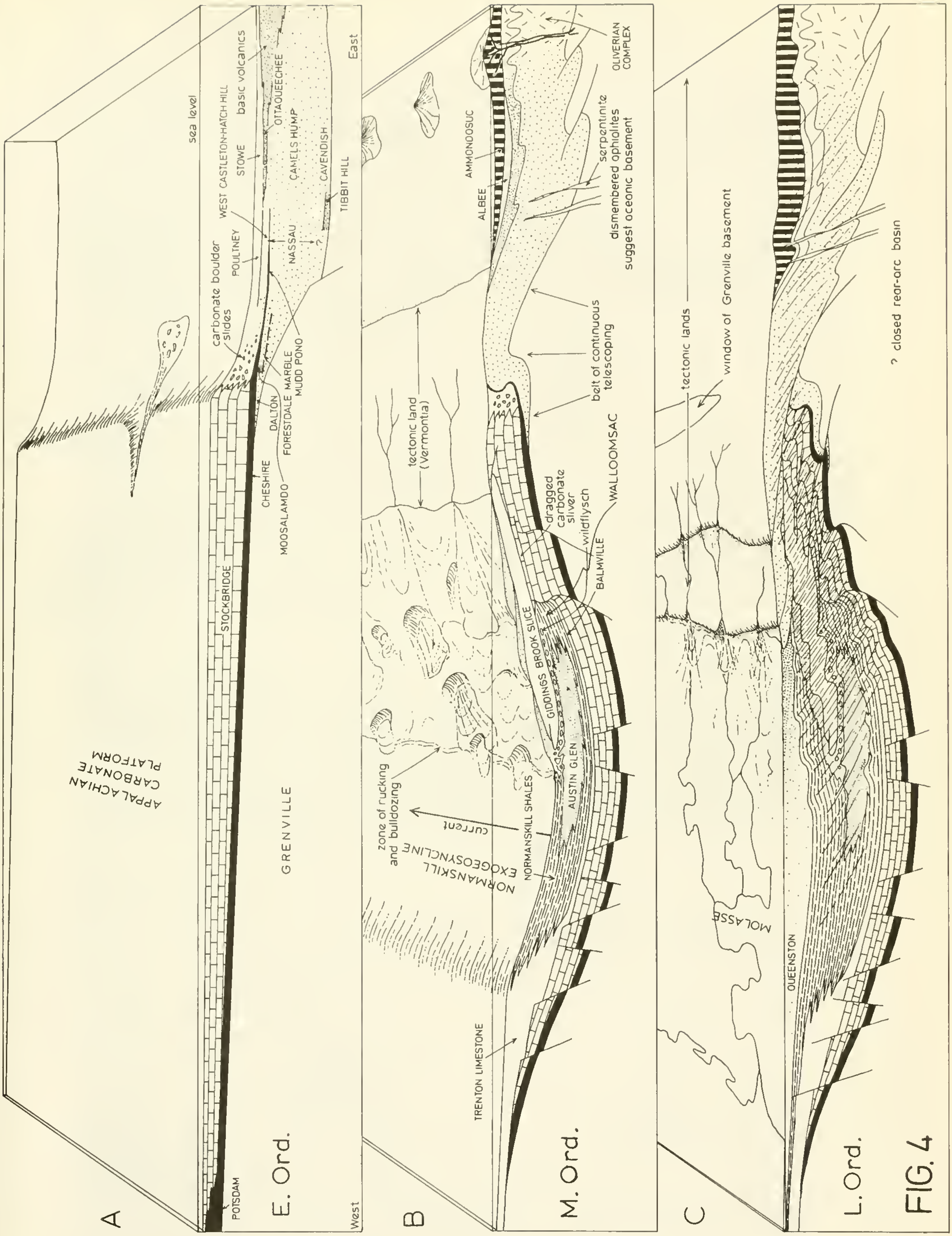

A
E. Ord.
continental shelf
continental rise
APPALACHIAN CARBONATE PLATFORM
sea level
GRENVILLE
POTSDAM
STOCKBRIDGE
CHESHIRE
MOOSALAMOO
DALTON
FORESTDALE MARBLE
MUDD PONO
carbonate boulder slides
POULTNEY
WEST CASTLETON-HATCH HILL
NASSAU
STOWE
OTTAQUEECHEE
CAMELS HUMP
CAVENDISH
TIBBIT HILL
basic volcanics
West
East
B
M. Ord.
TRENTON LIMESTONE
NORMANSKILL SHALES
NORMANSKILL EXOGEOSYNCLINE
current
zone of rucking and bulldozing
AUSTIN GLEN
GIDDINGS BROOK SLICE
BALMVILLE
wildflysch
dragged carbonate sliver
WALLOOMSAC
tectonic land (Vermontia)
belt of continuous telescoping
ALBEE
AMMONOOSUC
serpentinite
dismembered ophiolites suggest oceanic basement
OLIVERIAN COMPLEX
C
L. Ord.
MOLASSE
QUEENSTON
tectonic lands
window of Grenville basement
? closed rear-arc basin

FIG. 4

mechanism by which this relief inversion occurred is poorly understood. If one accepts the Ammonoosuc volcanic belt as a Paleozoic homolog of a present-day island arc, above a subduction zone, then the relief inversion was subduction-related. The models that have been suggested for this segment of the Appalachian Ocean continental margin include a subduction zone dipping west from a trench east of the Ammonoosuc belt and an east-dipping subduction zone from a trench between the continental margin and the Ammonoosuc belt. The latter model involves the collision of the continental margin with an Ammonoosuc island arc. Alternatively, the consuming plate margin relations may have been more complex. The Vermont ultramafic belt might represent the remains of a closed marginal basin (closed by east-dipping subduction) to the west of a west-dipping subduction zone as we have proposed for the equivalent region of the Appalachians in western Newfoundland (Bird and Dewey, 1970, Fig. 8; Dewey and Bird, 1971, Fig. 11). In any case, the timing of Ammonoosuc volcanicity, the timing of the Penobscot orogeny that includes deformation of the Bronson Hill anticlinorium in Late Cambrian - Early Ordovician times (see Rodgers, 1970), and the apparent westward-spreading vulcanicity and development of volcanogenic chert and flysch facies of Walloomsac time all indicate a westward migrating orogenic event that lead to the relief inversion of the continental margin. The initial phases of Taconic allochthon emplacement are only clear within the Taconic belt, as they were followed by later, intense deformation that greatly obscured any of the possible plate margin environments that existed oceanward in the "root zone". Continued detailed structural work in the Taconics and in the Ottaqueechee belt will provide data that will help refine our as yet schematic models of orogenic evolution of this segment of the Appalachians.

For the purposes of this guidebook the previous discussion is necessarily brief. We refer the reader to Bird and Dewey (1970) for a more comprehensive discussion of this proposed model of tectonic evolution of the Taconics, and of the theoretical plate margin relations that might have existed at that time. We have selected the particular trip stops, not only for their relevance to the plate model interpretations, but also because they all remain as still insufficiently understood aspects of Taconic geology that deserve continued future study.

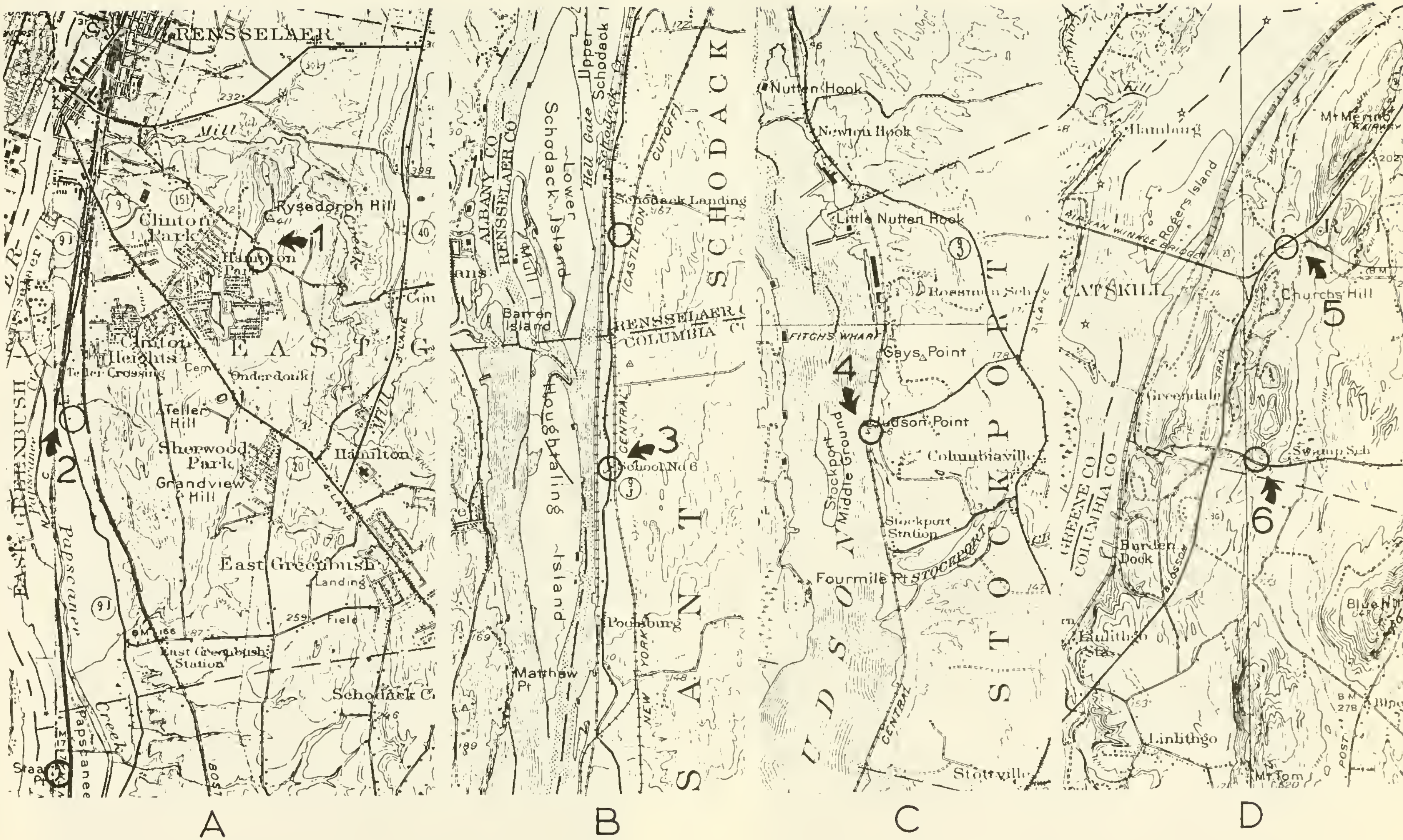


FIG. 5 TRIP LOCALITIES

1 mile

ROAD LOG

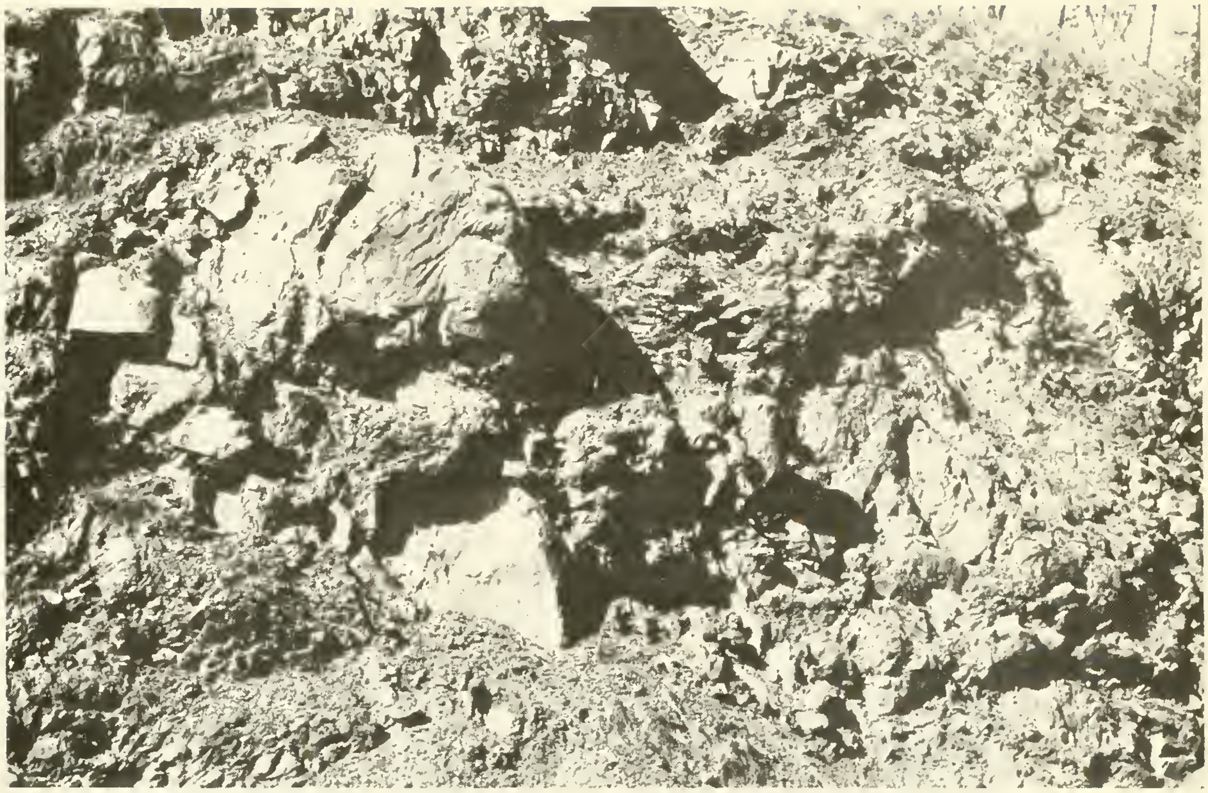
Start trip at road-cut on Route 151, just south of Rysedorph Hill, 1.7 miles southeast of the City Hall, Rensselaer, New York; see Fig. 5A. On the East Greenbush 7-1/2 minute Quadrangle the position is 4.0 inches east and 0.5 inches south of the northwest corner, and north boundary, along Red Mill Road between Couse and Rensselaer, just northeast of Clinton Park. Appropriate 7-1/2 minute quadrangles for the trip are Troy South, East Greenbush, Delmar, Ravena, Hudson North and Hudson South.

mileage

00.0 STOP 1 Wildflysch, Route 151 road-cut.
This exposure, of deformed Austin Glen graywacke and shale of the Normanskill, lies approximately 1.2 miles west of the front of the Giddings Brook slice. It is within the southern-most of three small hills; the famous Rysedorph Hill conglomerate (of Rysedorph Hill) occurs just to the north, and beyond that, approximately 0.4 miles, is Olcott Hill (see East Greenbush and Troy South 7-1/2 minute Quadrangles). All three of these hills are within the belt of wildflysch, called the Forbes Hill Conglomerate (Zen, 1961; see Fig. 1), which is attributed to a syntectonic bulldozing of flysch and incorporated, derived exotic blocks, in front of the advancing Giddings Brook slice. A general view of this "blocks-in-shale" (Berry, 1960) aspect is shown in Fig. 6A. A conceptual framework of the mechanism of formation of this mélange, or "wildflysch" (see Bird, 1969; p. 671), is illustrated in Figures 3 and 4B.

To evaluate the significance of the wildflysch we need to establish the stratigraphic/structural relations of the Giddings Brook slice that reveal its allochthonous nature. Figure 3 is meant to synthesize the regional stratigraphic and structural relations, especially those of syn-deposition of the Austin Glen Graywacke during the gravity sliding of the Giddings Brook slice. Essentially, the regional relations show that the gravity slide moved over Austin Glen sediment, commencing in Zone 12 time, somewhere east of the eastern-most extent of the carbonate miogeocline. To the east, at Whipstock Hill, near Bennington, Vermont (Potter, 1972), a wildflysch-like conglomerate overlies black slate of the Walloomsac containing Zone 12 graptolites (Zen and Bird, 1963, p. 45; Potter, 1972). On the west side of the Giddings Brook slice the only fossils that have been found in the shale matrix of the wildflysch are indigenous Zone 13 graptolites (Berry, 1962, p. 715), indicating that gravity sliding took place during an interval no longer than one graptolite zone.

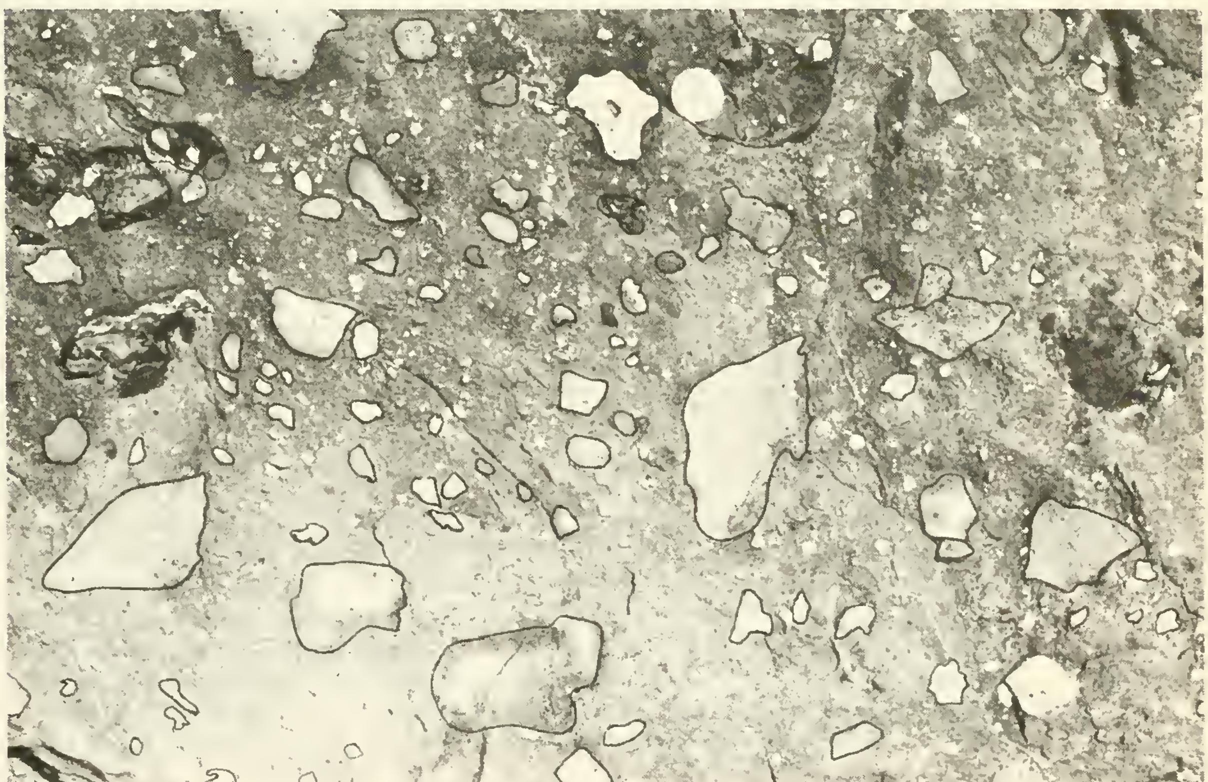
Most of the chips and blocks in the wildflysch are Austin Glen graywacke and shale. This lithic aspect is interpreted to be due to a predominance of "bulldozing" of the syndepositional flysch in front of the advancing Giddings Brook slice. Also, however, there are local occurrences of blocks of Giddings Brook slice



A. Route 151

FIG. 6 WILDFLYSCH

B. Staats Point



6 cm

strata, and carbonates that can be readily attributed to the carbonate sequence of the autochthon (we will examine some of those at Stop 2). The oldest-dated blocks are Elliptocephala (Lower Cambrian) fauna-bearing limestone, within the wildflysch exposed in the lower falls of the Moordener Kill. The Moordener locality is approximately 10 kilometers south of here, just east of the Brown Company paper mill which is indicated as a large group of buildings just north of Castleton-on-Hudson on the East Greenbush 7-1/2 minute Quadrangle. The locality is not a suitable stop for a large field-trip group; it is recommended, however, as a place to see the variety of lithic clasts that can be readily matched with Giddings Brook slice strata. The blocks include Zion Hill quartzite, West Castleton-Hatch Hill shale and limestone, Poultney shale, Mt. Merino chert, and the Rysedorph Hill Limestone Conglomerate. The "exotic blocks", i.e. not of the Normanskill shale/Austin Glen graywacke stratigraphic sequence, are thought to have been shed from the advancing Giddings Brook slice, into the frontal mélange, as illustrated in Fig. 4B.

This exposure is composed entirely of Normanskill/Austin Glen lithologies. The large blocks of graywacke "floating" in deformed shale are practically undeformed internally and were apparently completely lithified before deformation. Some of the smaller blocks, best seen in the rock-face on the south side of the road, have internal deformation that indicates they were semi-indurated at the time of formation of the block. In general, the spectrum of internal deformation features of the graywacke blocks of the wildflysch indicates there was a range of sediment consolidation, from soft to hard, within the autochthonous Austin Glen graywacke strata at the time of overriding and bulldozing by the advancing gravity slide. Although no precise measurements can yet be made to evaluate the deformational aspects of the zone of bulldozing, regional mapping shows a progressive decrease in intensity of deformation and size of blocks, to the west, away from the "front" of the Giddings Brook slice. For example, on the west side of the Hudson River, particularly around the Normanskill Creek locality just south of Albany (Albany 7-1/2 minute Quadrangle) one can see a gradual eastwardly increasing disruption of Austin Glen strata, from essentially flat-lying to the west, to bulldozed, in the creek-bed to the east, and also in exposures along the west bank of the Hudson River (Bird, 1969). The only exotic blocks known in the wildflysch west of the Hudson River are a few hills and knolls of Mt. Merino chert and shale such as Flint Mine Hill approximately 3 km southwest of Coxsackie, and Mettawee and Poultney(?) shale such as at the Barren Island, on the west shore approximately 1.6 km south of Coeymans. Exotic blocks older than the Mt. Merino of the Giddings Brook slice, and Trenton-age and older carbonates of the autochthon occur along the east bank of the Hudson (in the region of this field trip) and increase in frequency of occurrence eastward to the "front" of the Giddings Brook slice. This destribution of blocks suggests that at the time of gravity sliding the allochthon extended further west than its present limit, perhaps approximately as far west as

the western-most predominance of exotic (non-Austin Glen) blocks in the wildflysch.

The Rysedorph Hill Limestone Conglomerate occurs on the south side of Rysedorph Hill, about 375 meters north of the road-cut. (The exposures are small and difficult to show to a large group; we will not visit them). This famous lithology was studied in detail by Ruedemann because of its apparently unique fauna, ranging from Lower Cambrian to Trenton (see Zen, 1964a, p. 70, for summary of references and interpretations). Zen (1967) discussed the Rysedorph conglomerate in detail and showed that its occurrence, contained faunas, and its carbonate conglomerate character all indicate a syn-gravity sliding origin. We suggest that the Rysedorph and similar lithologies (Stop 6) may have originated in a fringing, carbonate-deposition environment on the leading edge of the Giddings Brook slice, where it was near sea-level (Bird and Zen, in preparation), and also perhaps on slide-block islands that slid off the front of the advancing allochthon into the wildflysch mélange environment. Perhaps Rysedorph Hill was such an island. In any case, the age-span of clasts of the Rysedorph conglomerate, from Lower Cambrian to Trenton, indicates a rock source for the clasts; the black shale (local) and Trenton-age fossil debris in the matrix of the conglomerate and its occurrence within the wildflysch matrix indicate a time of formation synchronous with gravity sliding.

The principal features to observe in the road-cut are the deformational aspects of the blocks and shale, and the size, shape, composition and distribution of the blocks. Note the "injection" of the black shale (mud) into some of the blocks, and the pervasive slatey cleavage of the matrix, and possible slatey cleavage within some of the blocks. (This cleavage is not the upright cleavage of Acadian-age seen throughout the region. It is thought to be either Middle or Late Ordovician age, formed during emplacement of the Giddings Brook slice, or the higher and later slices to the east). Elaterite (anthraxalite), a hardened hydrocarbon thought to be derived from petroleum, occurs in some of the graywacke blocks. Note the ferroan dolomitic (ankerite?) blocks, especially toward the west, on the south side of the road. These might be of a very local syn-gravity sliding facies (see Stop 6 discussion). Going from east to west, the stratigraphically upper portion of the outcrop, which has the usual black shale matrix, passes down-section to the west into green and grey shale having phacoidal slip-surfaces. These surfaces produce the structure known as argille scagliose, or scaley clay that is common in mélanges, both of subduction zones and wildflysch zones of exogeosynclines such as here. Note also the pronounced, laminated and annealed mylonite zones which cross-cut all lithologies of the mélange. These mylonites, which clearly post-date formation of the wildflysch, can be attributed to either the Late Ordovician emplacement of the high Taconic slices to the east which then also affected these rocks, or to the compressive events of the Acadian which superimposed deformation on the Taconic structures in Middle Devonian time.

0.3 Proceed 0.3 miles west along Route 151 and turn left on Sherwood Rd. (Rensselaer County #59).

1.0 Go 0.7 miles southwest to the T intersection of Sherwook Rd. and Routes 9 and 20, at Clinton Heights.

1.9 Turn right onto Routes 9 and 20 and proceed 0.9 miles northwest. Just past the bridge over the railroad, turn right at the sign indicating Castleton and Route 9J. Continue to bear right, going several hundred yards under the bridge and connecting with Route 9J where you bear left (south).

3.1 Follow Route 9J south. At 1.2 miles, outcrop of wildflysch on left (east) at base of hill with large tanks on top.

3.4 Continue 0.3 miles, past a gully, to the next outcrop on left, just opposite an old barn on the right, which is Stop 2 (Fig. 5A).

STOP 2 Wildflysch, Route 9J outcrop, 1.7 miles north of East Greenbush Station.

This road outcrop, recently exposed by the re-routing of Route 9J, is part of a belt that includes several excellent exposures along approximately 2 km of the New York Central Railroad tracks extending north from East Greenbush Station, about 0.3 km east of Route 9J (see East Greenbush 7-1/2 minute Quadrangle). Another similar exposure, having spectacular lithic aspects, occurs on the west side of Papscanee Island at Staats Point along the east shore of the Hudson River, approximately 2.4 km directly south southwest (on the border of the Delmar and East Greenbush 7-1/2 minute Quadrangles) of this stop. Further to the south is the previously mentioned Moordener Kill outcrop. Logistical constraints prevent us from visiting those exposures, although they bear on the discussion of this outcrop.

Most of the aspects of Stop 1 also apply to this stop. Here, however, the wildflysch is typical of that containing exotic clasts. Included are clasts of carbonates that can be referred to the carbonate miogeocline of the autochthon, and Mt. Merino chert, and Poultney(?) shale and limestone of the Giddings Brook slice. Some of the carbonate clasts contain fossils characteristic of Trenton faunules, and are either from the autochthon or the fringing reef environment invoked for the Rysedorph lithologies. A carbonate boulder, about 1.5 meters across, occurs near the central part of the outcrop. Note that the block is deformed internally; some of the carbonate is apparently reefal. Approximately 4 meters to the right (south) is a 10 cm cobble of white quartzite, with carbonate cement. This cobble is reminiscent of West Castleton-Hatch Hill sandstones. Another possible interpretation is that the cobble is of a local sand-carbonate, syn-gravity sliding facies that was disrupted and shed into the wildflysch. (A similar lithology fitting this interpretation occurs at Stop 6).

In addition to the variety of clasts and their shape and distribution, note the severe deformation of the shale matrix and the injection features which indicate soft, water-saturated-sediment conditions during deformation.

Figure 6B is of a portion of the Staats Point exposure (circle, lower left, Fig. 5A), and is included here to show better the aspects of the clasts in this facies of the wildflysch. At Staats Point there are also Trenton fossil-bearing carbonate clasts and one clast containing a Chazy-age faunule (D.W. Fisher, pers. comm., 1970) which lithically match the equivalent limestones in the autochthonous carbonate sequence. This is a significant occurrence because the clasts almost certainly came from the region of eastern carbonate belt (Stockbridge) and yet the Chazy-age fossils have not been reported from there. Ferroan-dolomitic carbonate blocks about 2 to 3 meters long, severely soft-rock-deformed Mt. Merino chert blocks, and chips and blocks of Poultney and Mettawee-like shale also are exposed at Staats Point. Along the contacts of the large carbonate blocks with the wildflysch shale matrix are quartz crystals up to 5 cm long that contain bubbles, now hardened, of what was a black fluid, probably the same as the elaterite seen at Stop 1. It is thought that the elaterite and the quartz crystals there and in many of the wildflysch outcrops, grew during development of the wildflysch under thermal conditions of about 100°-200°C, as the sediment was deformed and dewatered during overriding by the allochthon. No significant regional thermal event has been determined for these rocks or for the Lower Devonian rocks of the Heldebergs. Quartz and calcite mineralization also occurs in the Devonian rocks locally along fault surfaces.

Continue south on Route 9J.

5.1 Hays (Hayes) Rd. (Rensselaer County #58)

6.1 Staats Island Rd. (Staats Point)

8.1 Bridge over Moordener Kill

8.3 Road to Brown Co. paper mill (Moordener Kill wildflysch outcrop)

9.1 Junction with Route 150, in Castleton-on-Hudson

11.0 Berkshire Spur of N.Y.S. Thruway and N.Y. Central Railroad bridge

12.9 Intersection of Rensselaer County Rd. #2 and Route 9J, in Schodack Landing village

13.2 Park Inn Restaurant Bar, excellent wildflysch exposure in back yard; Berry's (1960) original "blocks-in-shale" outcrop (unnumbered circle, Fig. 5B).

14.0 Rensselaer County-Columbia County Line

14.9 Nearly one mile further is an upgrade and bend to the left, with a thin-bedded limestones-in-shale outcrop on the left (east) side of the road, just before the crest of the hill. There is an abandoned schoolhouse on the right side of the road. Just beyond the road-cut, on the left, is a dirt road with a cable barrier. Park at this turn-off, walk east about 75 meters, and then walk left (north) along the railroad tracks. The exposures are Stop 3 (Fig. 5B).

STOP 3 Nassau and West Castleton Formation rocks, exposures along the "Castleton cut-off" of the New York Central Railroad.

This stop (Ravena 7-1/2 minute Quadrangle), is the famous "Schodack Locality" of the Giddings Brook slice of the Taconic allochthon. The rocks exposed constitute some of the most significant of the Taconics in terms of the development of Taconic geology. The reader is referred to Theokritoff (1963) and Zen (1964a) for succinct accounts of previous work, regional correlations and the occurrences of the Lower Cambrian *Elliptocephala asaphoides* fauna. See Figure 2 for stratigraphic relations.

Exposures are poor in this region with the exception of those along the ridge overlooking the Hudson River, from here to Poolsburg, about one mile south. Mapping shows that this locality is along the front of the Giddings Brook slice; wildflysch crops out just below the cliff, about 150 meters west of the cut-off tracks. The wildflysch is severely deformed, with inclusions of Mt. Merino chert that apparently were soft when deformed. The overlying allochthonous rocks, as seen in the railroad cut, are only slightly deformed. The bedding dips generally eastward about 30 degrees, not quite parallel to that of the cliff exposures on the west side of Route 9J. The highest beds, at the southeast corner of the railroad-cut are not exposed elsewhere in the area. The following is a description of the section, downward from these uppermost beds which are referred to the West Castleton Formation. The transition to dominantly green-olive shale between units 7 and 6, is the transition zone into the uppermost part of the Nassau (Bull) Formation (Fig. 2). The section is measured downward through older rock in a direction almost due west (with dip correction) to the base of the cliff west of Route 9J.

		Thickness in feet
11.	Shale: black, finely fissile, with some thin siltstone and limestone beds; top not exposed	15+
10.	Massive, coarse sandstone	5
9.	Sandstone, shale, and sandy limestone	5
8.	Sandstone: one bed, in part conglomeratic with small limestone pebbles	3

7.	Shale alternating with sandstone and sandy limestone beds. One bed 13 feet below the top yielded inarticulate brachiopods, possibly including <u>Botsfordia caelata</u> (Units 11 through 7 constitute beds referred to by Zen [1964a, p. 75-76] in his discussion of the stratigraphic nomenclature of these rocks)	20
6.	Shale: dark-grey in upper part, grading to green shale near base	30
5.	Limestone pebble conglomerate in sandstone matrix; tapers northeastward and disappears. Better developed (4 feet) on west side of railroad cut. Some of the limestone pebbles aphanitic, blue-grey, weathering light-grey, identical with limestone of underlying unit; other pebbles crystalline, medium-grey, filled with <u>Elliptocephala</u> fragments	0-1.5
4.	Green shale, in lower part with nodules of aphanitic, tan-weathering limestone, forming transition to underlying unit	18
3.	Limestone: thin-bedded (1-3 inches), aphanitic, blue-grey, weathering light-grey, of very uniform lithology. Basal beds regularly bedded with thick shale partings, becoming nodular upward, in part brecciated near top. Forms extensive exposures in the cliff below the road for over 0.5 mile	14
2.	Green, silty argillite, weathering to tan slabs. In bottom 10 feet beds and lenses up to 6 inches thick of finely crystalline, light-grey limestone, yielding numerous specimens of inarticulate brachiopods and <u>Coleoloides</u>, and more rarely <u>Hyolithellus</u>. One large lens of limestone conglomerate occurs in this interval in the southernmost part of the exposures in the cliff between the road and the river shore. The matrix of the conglomerate is a grey limestone filled with quartz granules and pebbles up to a few millimeters in diameter, in places also containing numerous fragments of <u>Elliptocephala asaphoides</u> and opercula of <u>Hyolithellus</u>	45
1.	Green, finely fissile shale, forming sharp contact with overlying unit. Base nowhere exposed. At the base of the cliff at this locality, this unit rests on the wildflysch containing blocks of the Middle Ordovician Mt. Merino Chert.	60+
	Total thickness of beds exposed in the section	217+

Ford (1884) and Goldring (1943) reported fossils of the Elliptocephala asaphoides fauna from bedded limestone undoubtedly corresponding to unit 5 of the above section. Near its base and top, the conglomerate includes large slabs of crystalline limestone lying parallel to the bedding, and in limited exposures these may simulate beds in place. Careful examination showed that, in fact, all the limestone in this interval occurs in the form of pebbles and blocks. The characteristic, light- to dark-grey, crystalline limestone that contains innumerable fragments of Elliptocephala and other smaller, often better preserved, associated trilobites, seems to be rare as regularly bedded strata, even though it is common in similar conglomerates in Washington, Rensselaer, and northern Columbia counties. Lithically, the matrix of the conglomerate of unit 2 above is similar to some of these conglomerate boulders, even though the matrix contains more quartz pebbles than are usually found in these conglomerates. Ford (1884) reported Serrodiscus speciosus from limestone undoubtedly belonging to unit 3 of the above section on the basis of his descriptions, but the only fossil we (Bird and Rasetti) found in this interval is Hyolithellus.

Thus, the Elliptocephala fauna occurs in place in the lower portion of the Schodack Landing section and ranges through at least 50 feet of strata. Unfortunately, no diagnostic fossils were recovered from the higher units in the section; hence we do not know whether all the beds exposed here belong to the time span of the Elliptocephala fauna, or whether some portion of the beds is the time equivalent of strata carrying younger faunules elsewhere. Lithically, however, the uppermost beds correspond to the West Castleton Formation, that elsewhere (see next stop) also carries the Elliptocephala faunule.

There are several important aspects of this locality for the purposes of our trip. First, because of the faunules found in this section and the fact that the black-green boundary between the West Castleton and Nassau beds is homotaxial throughout the Giddings Brook slice (Theokritoff, 1963) we can take this section as being representative of the fossiliferous base of the Cambrian, and representative of the mapping boundary between the West Castleton and Nassau Formation. The section corresponds in age to the Olenellus-bearing Cheshire Quartzite of the miogeocline (see Fig. 2). Elsewhere, the Mudd Pond Quartzite (see Zen, 1964a) occurs within equivalents of this section. It is a rounded-sand quartzite that is sporadically developed and although not present here, is comparable to unit 10 of the measured section. The Mudd Pond may contain the lowest occurrence of the Olenellid fauna, at Diamond Rock, Troy, New York (see Bird and Rasetti, 1968), because there the quartzite is within Nassau beds that appear, by mapping, to be lower in the section than the lowest fossils found here (unit 2); no Elliptocephala faunas have been found stratigraphically lower.

Second, these exposures are in an overthrust relation with the Normanskill strata of the Hudson Valley. At the base of the cliff

is wildflysch which is of the regional belt (see Fig. 1) that everywhere has Zone 13 (Berry, 1960) graptolites. Going away from this contact to anywhere in the Taconic region, the facies of this locality is always found to be overthrust or allochthonous, and can only be tied to the Normanskill Cheshire sequence of the sutochthon via the age-correlation of the Olenellid faunas and not through a conformity of strata. The relations lead to a third aspect, that of the depositional environments that can be determined from the structures and petrology of the strata.

It has been suggested (Bird and Rasetti, 1968; Bird and Dewey, 1970) that these strata accumulated in a rise environment, off-shelf from the early-developing miogeocline. The Mudd Pond Quartzite may represent a sand-flush from the shelf of Cheshire sediment. The bedded limestone, such as in units 3 and 4, are apparently carbonate mud turbidites; the conglomerates such as unit 5, may be slump-derived from a bank edge. Commonly, these limestone pebble conglomerates have at least three varieties of limestone clasts, and a rounded-sand matrix, all entirely devoid of a shale component comparable to the surrounding shale beds. Conversely the thin-bedded limestone section (unit 3) appears to have broken up into conglomerate and is an indication of syn-depositional slope-instability. Figure 7A, of this unit, shows a limestone slab penetrating another, which is clearly criteria of both soft sediment and different states of induration conditions at the time of formation of the individual slabs. Also, there are a variety of current marks on bedding surfaces and load casts, cross-lamination, and graded bedding, both within regularly bedded strata and in the clasts of the polymict conglomerates, as well as in the thin, boudinaged to chaotically broken calcareous layers. The argillaceous rocks are bioturbated in many places. Various forms of deformed laminae occur within the regularly bedded strata. All these features indicate bottom current activity, deposition from turbidity currents and sediment slumping and slope instability in the environment of deposition of the strata. The carbonate and sand fractions were apparently derived from the correlative sand and carbonate terrains of the miogeocline. Therefore, this facies is taken to be of the early continental rise of the opening (post-graben) stages of the Northern Appalachian-Atlantic Ocean.

Continue south on Route 9J.

17.9 New York Central Railroad overpass

19.6 Village of Stuyvesant and intersection with Route 398

22.2 Intersection with Stuyvesant Falls Rd.-Ferry Rd., in Nutten Hook (Newton Hook)

24.7 At the end of Route 9J, intersection with Route 9, turn right (west) onto deadend, paved road to Judson Point

5 cm

B-1

A. Castleton cutoff

FIG. 7 CAMBRIAN CONGLOMERATES

B. Judson Point

10 cm

25.7 Descend hill and winding road and bear right at Southers Rd., about half way down the hill

25.9 Road ends at New York Central Railroad main line tracks. Exposures from old building foundation on south of road, over the hill to the railroad are Stop 4 (Fig. 5C).

STOP 4 Nassau and West Castleton-Hatch Hill Formation rocks, Judson Point (Fig. 5C).

This stop is included because strata here extend up-section from those at the Castleton cut-off, and the exposure is perhaps the best of the entire Taconic region in which to study in detail sedimentologic aspects of portions of the proposed continental rise facies. Mapping indicates that this hill is a detached block of the Giddings Brook slice, "floating" in the Forbes Hill Conglomerate or wildflysch zone, in front of the Giddings Brook slice (Fig. 1). The sequence consists of strata ranging from the upper part of the Nassau Formation into undifferentiated West Castleton-Hatch Hill Formation. Dark-grey shale outcrops in the cliff just south and east of the old building foundation pass westward into alternating sandstone, shale, limestone and conglomerate beds for several hundred feet in the cliff-face of the railroad cut. Detailed examination of the sequence shows that there are no significant tectonic breaks or unconformities present. The rocks are remarkably undeformed compared to the wildflysch. The bedding dips uniformly east at 40 degrees; sedimentary structures and unquestionable fauna evidence show the entire sequence to be inverted. The following is a description of the strata in descending order from unit 27 (along the railroad tracks) which is the youngest but geometrically lowest part of the sequence.

	Thickness Feet	Thickness Inches
27. Shale: dark-grey, with some thin-bedded limestone. (Younger beds concealed under railroad tracks)	6+	
26. Sandstone: one bed	2	
25. Shale and Sandstone	14	
24. Shale	2	
23. Shale and sandstone: the latter partly thick-bedded	11	
22. Sandstone: massive, tan-weathering, two beds	6	
21. Shale and sandstone	3	

20.	Shale and sandstone: at top lens of rusty-weathering polymict conglomerate 0 to 1 foot thick	9	
19.	Shale: dark-grey, fissile		7
18.	Shale and interbedded lenses of grey, light-weathering limestone		9
17.	Limestone: one bed, light-grey, sandy, in part rusty weathering, contains: *Pegmatreta* cf. *ophirensis* (Walcott) *Baltagnostus* sp. *Bathyuriscus* sp. *Centropleura* sp. *Hypagnostus parvifrons* (Linnarsson) This fauna is of the Late Middle Cambrian BOLASPIDELLA (North American) faunizone.		3-5
16.	Shale alternating with rusty-weathering siltstone	1	1
15.	Polymict conglomerate, including flat limestone pebbles of different lithic types and some siltstone pebbles, in a coarse, sandy, and calcareous, rusty-weathering matrix. Tapers and disappears southward (Fig. 7B)	0-2	
14.	Shale, siltstone, and limestone: most of the beds rusty-weathering	2	4
13.	Limestone: fine-grained, dark-grey, weathering light-grey, in thin, nodular beds alternating with black shale	1	6
12.	Limestone: silty, on bed		4
11.	Shale and sandstone	2	
10.	Conglomerate of calcareous pebbles in sandy, rusty-weathering matrix. Tapers and disappears northward	0-2	
9.	Sandstone: rusty-weathering, one bed	2	6
8.	Shale and thin-bedded sandstone		3
7.	Sandstone: coarse, thick-bedded, rusty-weathering	8	
6.	Shale and sandstone	3	

5.	Sandstone: coarse, rusty-weathering, with some shale. Unit forms the crest of the railroad cut at the north end	9	6
4.	Sandstone: coarse, thick-bedded, rusty-weathering	3	
3.	Shale and thin-bedded, rusty-weathering sandstone	11	6
2.	Siltstone, weathering to irregular fragments	6	
1.	Shale: dark-grey, in upper part finely fissile, splitting perfectly along bedding planes, grading downward, stratigraphically, to more silty or calcareous shale that breaks with conchoidal fracture approximately parallel to bedding. Continuously exposed for 80 feet, the exposures of lower beds somewhat discontinuous but certainly still representing a sedimentary sequence; about 100 feet below stratigraphic top of unit grades to coarser shale and siltstone, and becomes progressively more olive-green-colored	130+	

Fossils collected in this unit at 30 feet and 38 feet below stratigraphic top in lenses of more calcareous shale, and 80 feet below top in an interval of green shale. Same fauna in all collections:

- Hyolithellus sp.
- Hyolithes sp.
- Obolella sp.
- Atops trilineatus (Emmons)
- Elliptocephala asaphoides Emmons
- Serrodiscus speciosus (Ford)
- Rimouskia typica Resser

This fauna is of the Lower Cambrian BONNIA-OLENELLUS (North American) faunizone.

The faunal evidence shows that the rocks of this section range from the upper Lower Cambrian to the upper Middle Cambrian; this is the only exposure of the Taconic sequence known to show the transition from Lower to Middle Cambrian fossiliferous strata. The thickness, representing what must be a long time interval, is surprisingly small, considering that the youngest Middle Cambrian faunule in the area, the Centropleura faunule of unit 17, occurs only 85 feet above the Elliptocephala asaphoides faunules of unit 1 which represents the oldest known Early Cambrian fauna of the Taconic sequence.

Approximately 370 meters to the north-northeast, there is an

exposure of fine-grained, thin-bedded, grey limestone in grey shale at the shoreline and in the cliff, on the east side of the pond which is separated from the Hudson River by the New York Central Railroad tracks. Very small, rare, immature trilobites were found in a slightly granular portion of a 5 cm thick limestone bed in the shoreline exposure. The fossils include Richardsonella sp. and Theodenisia sp. which indicate a Late Cambrian, probably Trempealeauian age. The rocks are severely deformed and are, apparently, of another block in the wildflysch. The lithology resembles lower Poultney units, and are similar to exposures on the west shore of the north hill at Nutten Hook, and under the Columbiaville bridge.

Figure 7B is of the conglomerate in unit 15 of the described section. Note the variety of lithologies constituting the clasts in this bed. Some of the clasts are clearly of shallow-water, carbonate-mud environments; some of the sand grains are practically spherical and almost certainly of a beach or aeolian origin. Note that this conglomerate and similar ones further south in the outcrop lack any significant shaley fraction or shale clasts, which suggests loaded or perhaps fluxoturbiditic transport conditions. Many of the beds are composed of cross-laminated sand in a carbonate matrix, and are graded, as typical turbidites.

Bedding-surface strucutres, including "worm tracks", characteristic of turbidites are common on these and other thin calcareous and sandy beds throughout the upper half of the section. Note also that some of the thin-bedded, turbiditic carbonates appear to have reacted with shale, perhaps going from an initial calcitic composition upon deposition, to ferroan dolomitic composition by diagenetic reaction with the more iron-rich shale (mud) of the deeper-water, rise environment. The massive sandstones, containing carbonate clasts, might have been sand-flows from the shelf region.

In addition to the variety of sedimentologic features at this locality, the faunas also indicate these sediments accumulated in an off-shelf, starved-rise environment. As previously mentioned, the section from the Lower Cambrian faunules to the Upper Middle Cambrian is only about 100 feet (30 meters) thick. This is a surprisingly thin amount of sediment accumulation for such a long period of time. It corresponds to several thousand feet of the correlative Cheshire-Monkton-Winooski strata of the autochthonous shelf sequence. Also, the Centropleura-bearing bed (unit 17) is particularly interesting sedimentologically. These Middle Cambrian fossils were found by carefully searching with a hand-lens the weathered surfaces of the beds. The fossils, mostly complete immature forms, are within the grading sequence of the grains of the bed! Apparently the fossils were sorted along with the rest of the sediment during density-current transport from the carbonate environment source, to deposition as a turbidite in this shale environment.

Bird and Theokritoff (1967) originally suggested that most of the faunas of the Cambrian of the Taconic sequence were indigenous to the carbonate deposition environments of the shelf, or miogeocline, and its edge, and that all of the shelly faunas, such as here and at Stop 3, were transported from the shelf via density currents or slides and slumps of sediment (e.g. Elliptocephala-fauna bearing clasts in conglomerate), into the deeper-water mud environments. However, here at Judson Point complete specimens of Serrodiscus speciosus found in unit 1 of the measured section show no evidence of having been transported. This trilobite and Atops trilineatus (Rasetti, 1967), both of the Elliptocephala asaphoides fauna, may have adapted to the rise environment and therefore may actually be indigenous (see Theokritoff, 1968 for detailed discussion).

Please don't needlessly hammer this outcrop. Practically all the features can be best seen in the weathered surfaces, and as is true in most of the Taconics, good outcrops are hard to come by!

27.1 Turn around and go back to the intersection of Routes 9J and 9. Turn right on Route 9, going south toward Hudson, N.Y.

28.2 Columbiaville Bridge, over Kinderhook Creek

30.3 Stottville intersection, continue south on Route 9

33.1 City of Hudson line

33.3 Turn right on Routes 9 and 23B west

33.8 Junction of Routes 9, 9G, 23B west

33.9 Bear right (west) on Columbia St. and follow signs for Rip Van Winkle Bridge, along Routes 9G and 23B.

34.5 Turn left (south) on South 3rd Street, Routes 9G and 23B

35.2 Railroad tracks. Hill ahead and to right (southwest) is Mt. Merino.

37.4 End of Route 23B, junction of Routes 9G and 23. Outcrop on south side of Route 23 is Stop 5

STOP 5 Mt. Merino Chert and Shale, Mt. Merino (Fig. 5D)

The formal name Mount Merino Chert and Shale was proposed by Ruedemann (1942, p. 90) for these rocks. Beds exposed at the north end of Mt. Merino are apparently the type-locality. The Mt. Merino is composed of interbedded shale, siliceous shale and argillite, and green and black chert. It overlies the Indian River Slate of the northern Taconic region (equivalent to Hudson Red and Green Slate of Dale, 1899). Ruedemann and Wilson (1936)

studied the "Normanskill cherts", and the older "Deepkill cherts", and concluded that the chert beds were originally accumulations of colloidal silica derived from submarine or continental volcanic activity. They describe radiolarian forms that occur in the chert. Ruedemann (1942), and Ruedemann and Wilson (1936) found that most of the Mt. Merino rocks occur in a NNE-trending belt of the western Taconics (Giddings Brook slice), east of a belt of graywacke and shale (Austin Glen facies). Ruedemann (1942) pointed out that both belts contain graptolite faunas characteristic of the "Lower Dicellograptus Zone", but the two rock assemblages are not interbedded. Ruedemann's only evidence to suggest they might be interbedded are a few exposures of black siliceous shale containing sparse, thin, sandy layers similar to Austin Glen graywacke (Ruedemann, 1942, p. 88). These outcrops, and the faunal evidence are probably what led Ruedemann (1942, p. 88) to construct his Normanskill Formation, being somprised of chert and black shale of the Mt. Merino exposures and the Austin Glen strata. Following this, Berry (1962) studied in detail the faunal assemblages of the various "Normanskill" strata including the Austin Glen and Mount Merino and found graptolite assemblages corresponding with his graptolite zones of the Mariavillas Chert and Shale of the Marathon, Texas, region (Berry, 1960). Berry (1962) also found that many of Ruedemann's "Deepkill cherts" are not Mt. Merino age but are correlative with upper portions of the Poultney Formation of the northern Taconic region, which ranges from Early to Middle Ordovician. The youngest Poultney is Zone 12 in age (unit 4 of Zen's Mt. Hamilton Group, see Zen, 1964a), correlative with the Mt. Merino. Berry then divided the "Normanskill" into four members, equivalent to the Indian River red slate and chert, the Normanskill shale, the Mt. Merino shale and chert, and the Austin Glen graywacke and shale. Bird (1969) discussed these relations in detail and showed that most likely the Indian River-Mt. Merino rocks are the uppermost facies of the Poultney Formation, are entirely allochthonous, and are not a stratigraphic part of the Normanskill rocks (see Fig. 2) of the exogeosyncline into which the Giddings Brook slice was emplaced. More recent mapping, particularly in this region, supports this and shows that the Indian River-Mt. Merino shale and chert facies is in stratigraphic continuity with underlying Poultney (Stuyvesant) strata. Conversely, in the northern Taconic region Zone 12 graptolites have been found in Pawlet Formation graywacke which is locally conformable above the Mt. Merino. Zen (1961) also showed that the Pawlet is unconformable on older rocks of the Giddings Brook slice and pointed out that the Pawlet is equivalent with the Austin Glen facies (Zen, 1964a). Therefore, the graywacke facies spans Zone 12 and 13 time, and its regional distribution and stratigraphic relations show that the facies was syn-tectonic with the emplacement of the Giddings Brook slice (Fig. 3); the Indian River-Mt. Merino chert facies apparently predates the conversion of the shelf-rise couple into the Giddings Brook allochthon/Normanskill exogeosyncline complex.

Essentially, then, whereas Berry's (1962) Members 1, 2 and 3 of his defined Normanskill belong to the Climacograptus bicornis Zone (Zone 12 of the Marathon sequence), only his defined Zone 13 (Orthograptus truncatus var. intermedius Zone) graptolite assemblage

has been found in the wildflysch matrix west of the Giddings Brook slice, and in the autochthonous Austin Glen facies. Mapping shows that Members 1, 2 and 3 are either within the Giddings Brook slice or as blocks in the wildflysch. Furthermore, mapping in this region shows Mount Merino and Mount Tom (Mt. Thomas) to the south, to be large blocks within the wildflysch, either as erosional remanents or detached pieces of the Giddings Brook slice (Bird, 1969). It seems clear from these relations that, although they are intimately associated, both with the Giddings Brook slice and in the wildflysch, the Mt. Merino and Austin Glen facies were stratigraphically associated only during initial development of the Giddings Brook slice, in the region off the continental rise.

On the basis of graptolite chronology (Berry, 1963; Harwood and Berry, 1967) the Indian River-Mt. Merino rocks, and the early flysch facies (Pawlet, and some Walloomac) are synchronous with the Ammonoosuc "spilite" facies, east of the Berkshire-Green Mountain anticlinorium. This relation led Bird (1969) and Bird and Dewey (1970) to propose that the sub-Trenton/Black River unconformity of the shelf sequence developed as a respense to initiation of Ammonoosuc volcanism, and that in the off-shelf region the Indian River-Mt. Merino chert facies accumulated from colloidal silica derived from the vulcanism. Ruedemann (1942) concluded that Mt. Merino radiolaria indicate that the cherts accumulated in an abyssal environment, perhaps about 4000 meters deep. Zen (1967, p. 47) proposed such marine conditions for the Poultney and suggested the Timor Trough, adjacent to the Sahoel shelf of Northwestern Australia as a modern analog. Bird and Dewey (1970) proposed that the Ammonoosuc volcanics were of an island arc behind a subduction zone and that the Pawlet-Austin Glen flysch migration which "overwhelmed" the chert environments, was a result of westward-spreading vulcanicity, preceeding and synchronous with the evolution of the Giddings Brook slice (Fig. 4B).

Petrographically, the cherts of the Mt. Merino are found to be complex assemblages of thin laminae enclosing lenses of distinct clastic fragments, rare authigenic minerals, carbonate euhedra, carbonaceous shreds and radiolaria and radiolaria-like forms. Usually the lenses are deformed into planar, streaked-out shapes that indicate pre-lithification, diagenetic alteration of original bedding geometry. The varieties of the laminae include aggregates of non-clastic quartz micro-crystals, less than 0.02 mm and interspersed in brown isotopic material (1.55 RI); fine mosaic quartz less than 0.02 mm; coarse mosaic quartz greater than 0.03 mm; sparse, irregularly bounded, interlocking quartz crystals less than 0.02 mm, all elongate parallel to bedding, forming a "shredded quartz" network; numerous, irregularly bounded, interlocking crystals, less than 0.02 mm, all elongate parallel to bedding, forming a "spongy quartz" network; and felted masses of quartz which form a "semi-continuous quartz" network. Subangular grains of quartz and feldspar less than 0.02 mm in diameter are common; twinned sodic plagioclase is common while microcline and zoned plagioclase are rare. The boundaries of most of the grains are corroded. These grains are definitely clastic; they constitute

less than 5 percent of most siliceous laminae. Other clastic grains found include chlorite, augite, hypersthene, garnet, biotite, schlorite, microcline, apatite, rutile, barite, all usually dispersed throughout the laminae. Some laminae contain euhedral to anhedral carbonate grains, 0.02 mm or less, dispersed in the chert. These include calcitic, dolomitic, and ankeritic compositions; the grains are perhaps both authigenic and clastic. Definite clastic and authigenic laminae of carbonate occur. Very fine-grained carbonate occurs in many laminae that are characterized by abundant spherules and rods of chamosite; the groundmass of these laminae is mosaic quartz. Siderite is rare, and occurs as spherulitic nodules surrounded by aggregates of euhedral pyrite and corbonate in a quartz matrix. Pyrite occurs in most laminae; granular aggregates are common in some laminae. The pyrite has selectively replaced larger carbonate euhedra; and some spherulitic forms.

The thin laminae that are outlined by various amounts of the fine-grained clastics (clay, quartz, feldspar), carbonates and sulphides within the textural distinct groundmass of quartz aggregates, both mosaic and felted, are almost certainly primary in origin. The mosaic quartz laminae probably formed from precipitated colloidal silica, while the felted network probably formed beneath or between mosaic quartz laminae, as a result of retarded silica diffusion during early diagenesis (possibly the silica had been adsorbed by clay minerals within the laminae during initial deposition). The authigenic minerals may represent stable mineral assemblages formed as a result of mineral equilibration during deposition, or almost certainly, during diagenesis.

Radiolaria are relatively scarce in the Mt. Merino chert. This, and the petrographic aspects mentioned, indicate that the beds did not accumulate as "radiolarian ooze". Rather it is thought, from both stratigraphic and petrographic aspects, that the beds accumulated relatively rapidly compared with computed rates of deposition of radiolarian ooze (0.5 to 1.0 cm/1000 years, equal to 0.17 to 0.33 cm of dry sediment, Strakhov, 1962). Radiolaria usually occur as lenses of radiolarian tests, within radiolaria-free beds, suggesting that radiolaria may have been abundant only near the ocean surface or at moderate depths. (Some of the lenses are xenolithic in the chert and may have agglomerated at or near the sea-surface and then fallen as clasts into the colloidal silica on the sea-floor).

This outcrop is fairly typical of the Mt. Merino Chert and Shale in this region. The chert beds are usually less than 0.6 meter thick, and most commonly, are less than 10 cm. The shaley partings are usually less than 5 cm. The thickest known sequence, about 38 meters, is at Fly Summit, Washington County. Berry (1962) estimated the total thickness of the Mt. Merino units to be 150 to 230 meters thick. More recent mapping shows the thickness to be less than 75 meters in the Giddings Brook slice. The color here is also typical.

Zone 12 graptolites have been found in this outcrop (identification by Berry, pers. comm. to Bird, 1969).

Note the fault contact of the chert beds with the Austin Glen graywacke and shale strata, about 2/3rds west from the eastern end of the outcrop. The graywacke and shale may also be a block in the wildflysch, which outcrops further west along the highway, near the junction of Routes 23 and 9J, before the Rip Van Winkle Bridge. Also, see the exposure just to the north on 9J; Stop 2 of trip A1, this guidebook.

37.8 Continue southwest on Routes 23 and 9G and turn left (south) at 9G and 23 intersection. Churchs Hill on left (south).

39.2 Follow Route 9G to intersection with Columbia County Road #13

39.4 Turn left onto #13 and proceed east to road-cut, which is Stop 6 (Fig. 5D)

STOP 6 Burden iron ore, south end of Cedar Hill

This outcrop is within the southern-most of the several ridges extending south from Churchs Hill, just south of Stop 5 (see Hudson South 7-1/2 minute Quadrangle). On the basis of detailed mapping it is thought that the several hills of the area are either on-strike separate blocks, or of the same block that constitutes Mt. Merino, within the wildflysch. The outcrop is comprised of ferruginous quartzite, and carbonate beds and argillite, part of the sequence of sideritic and limonitic Burden iron-ore strata that were mined in these hills, and especially at Mt. Tom to the south, in the late 1800's.

Ruedemann (1931, 1942) discussed in detail the origin of the Burden iron ore. He indicated (1931) that the ore was indigenous to Austin Glen strata ("Normanskill grit"), resulting from the contemporaneous alteration of magnetite sand with the calcareous matrix of the "grit", to form siderite; the siderite was then altered to limonite. Magnetite (and pyrite) have not been found in Burden iron ore-bearing strata. Later, he (1942) indicated the Burden iron occurs between the "Nassau" and "Schodack Formations" (this would correspond to the lower and middle beds of the section at Stop 4, Judson Point). He suggested that the rocks might have been structurally dislocated from the Copake region 22 km to the east, where similar ores occur. Ruedemann (1942) cites field and petrographic evidence that shows that the Burden ores were deposited as sediments, contemporaneously with the bounding strata. The ores are restricted to the crests of the various local hills.

We have selected this stop because the Burden iron ore remains as one of the more interesting unsolved geologic problems of the Taconics. Our preliminary results of the study of these

rocks suggests an alternative explanation of origin for the iron ores. It is possible that the ferruginous quartzite and carbonate beds, and local carbonate and sand conglomerate, originated in a fringing reef and shore environment where the upper portions of the supposed slide-blocks in the wildflysch were exposed as islands in the Normanskill sea. Such an environment would produce quartz and carbonate sand, and conglomerates, along the slopes of the blocks. Possibly, the ferruginous carbonates formed by reaction of carbonate precipitates with the iron-rich, deep-water facies sediments of the allochthonous blocks. It is interesting to note that the ferruginous carbonate blocks seen in the wildflysch at Stop 1 are very similar to carbonate beds exposed here. Also, the Austin Glen facies is quite carbonate-rich in the region west of these hills. Perhaps some of the enigmatic carbonate conglomerates of the wildflysch, such as the Rysedorph, also had their origins in carbonate-reef and shore environments of moving islands of separated blocks in front of, and on tectonic salients along the leading edge of the advancing Giddings Brook slice.

End of road log.

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Fold-Thrust Tectonism in the Southern Berkshire Massif, Connecticut and Massachusetts

by

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Introduction

The western front of the Berkshire massif consists of imbricate thrust slices of Precambrian and lower Paleozoic rocks transported westward over Cambrian and Ordovician rocks of the carbonate platform sequence. Thrusting was accompanied by large and small scale recumbent folding, attenuation and transposition along the overturned limbs of the recumbent folds, and the formation of cataclastic foliation, now recrystallized, in the thrust zones. The folding, thrusting, and cataclasis define a tectonic style for the Berkshire massif that closely resembles the tectonic style of the Reading Prong described by Drake (1970) and one that may be characteristic of most, if not all, of the Precambrian massifs along the backbone of the Appalachian orogen. It seems appropriate, therefore, to examine the rocks in some of the thrust zones to see the textures and structures that make up the fold-thrust fabric and the fold-thrust style of tectonism.

Although the fold-thrust style of deformation persists throughout the Berkshire massif, there are some significant variations in the structures and lithologic sequences along its length. North of the Massachusetts-Connecticut State line (fig. 1), the thrust slices are stacked up, shinglelike, to the north, so that successively higher tectonic levels and, perhaps consequently, more brittle structures are exposed northward. South of the State line, the thrust slices stack up to the south and are overlapped at the southe end of the massif by the parautochthonous Waramaug Formation of Gate (1952, 1965) which traces into the Hoosac Schist (Lower Cambrian or older) east of the massif. The Waramaug Formation rests structurall above a sequence of feldspathic granulites, volcanic rocks, and garn sillimanite schist, referred to collectively as the Canaan Mountain Schist by Rodgers and others (1959). Lithologically, the rocks of t Canaan Mountain Schist appear to be a transitional facies between th Dalton Formation, the basal clastic unit of the platform sequence to the west, and the Hoosac Schist and possibly the Rowe Schist, the lower units of the basin sequence east of the massif. This trip offers the opportunity to compare the rocks of the Dalton Formation with some of those of the Canaan Mountain Schist and to see the contact relationships between the platform sequence and the Canaan Mountain Schist.

Most of the stops are on private property; please respect it. Also, most of the geologic features, particularly the cataclastic features, show up best on the weathered surface. There is no need to quarry the outcrops which don't yield readily to the standard size geologic hammer anyway. The optimum cost-benefit ratio for hand-specimen collecting is associated with the talus blocks.

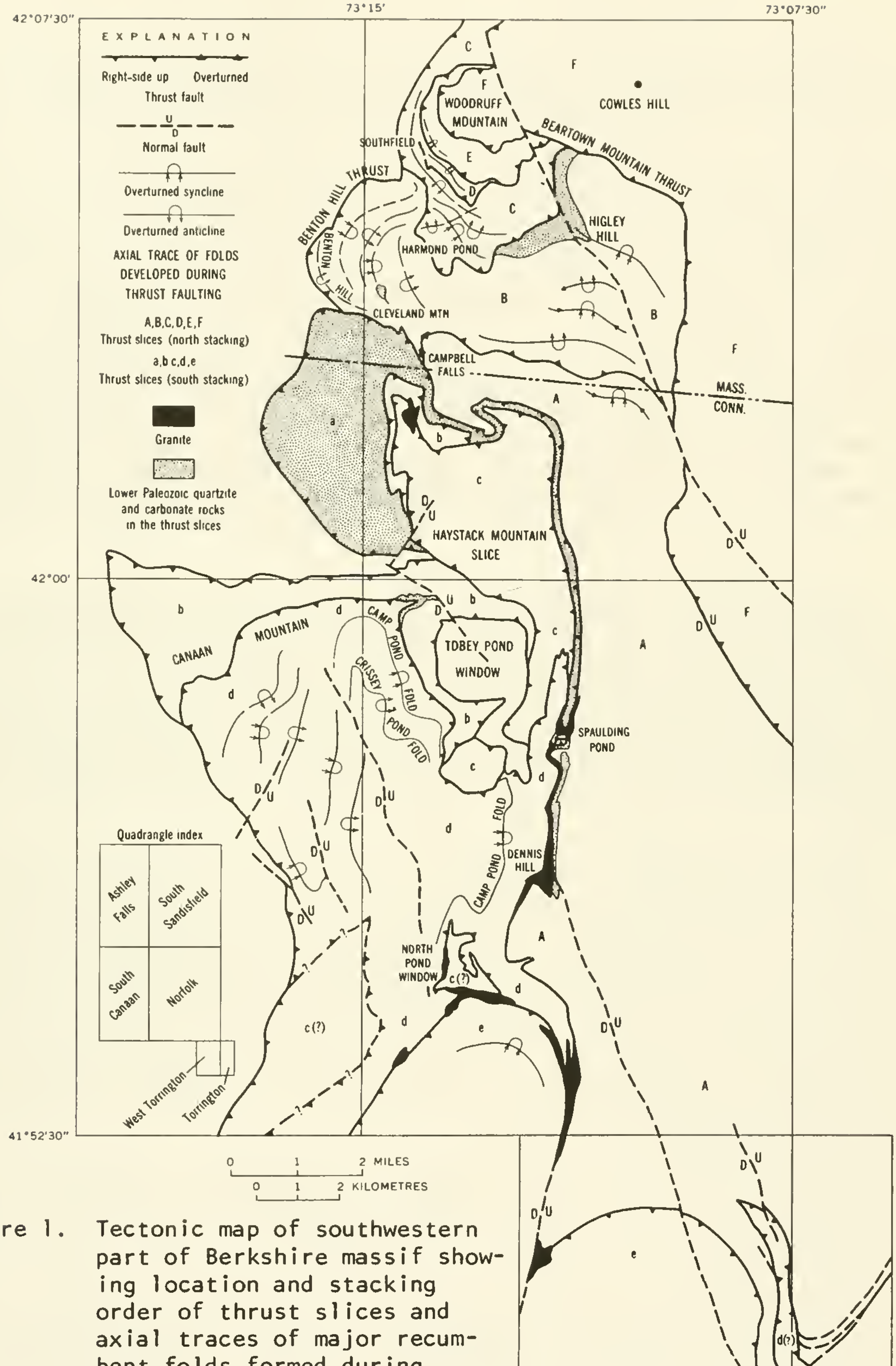


Figure 1. Tectonic map of southwestern part of Berkshire massif showing location and stacking order of thrust slices and axial traces of major recumbent folds formed during thrusting.

Stratigraphy of the Precambrian Rocks

A thin but persistent calc-silicate marker unit has been the key to unraveling the stratigraphy and the complex recumbent folding and thrusting in the southern part of the Berkshire massif. The calc-silicate unit is part of a sedimentary-volcanic paragneiss sequence that was intruded by granite, regionally metamorphosed, and complexly folded about east-trending axial traces about 1 billion years ago (Ratcliffe and Zartman, 1971). Although the granite gneiss is locally discordant and clearly younger than the paragneiss sequence, it is broadly conformable to the other units and will be treated as a stratigraphic unit here. The normal stratigraphic sequence is arbitrarily taken as the succession of units away from the Dalton Formation on Higley Hill (fig. 2), where the Precambrian-Paleozoic boundary appears to be an angular unconformity uncomplicated by thrust faulting. This assumes the Dalton was deposited on right-side-up Precambrian rocks at that point--an assumption that cannot be proved, but one of little consequence with regard to understanding the Paleozoic deformations. It is important to remember, however, that the Precambrian stratigraphy has been determined concurrent with an in respect to Paleozoic structures.

At Higley Hill, the Dalton rests unconformably on biotite-quartz-plagioclase gneiss. Near its base, the gneiss tends to be dark gray and well layered and commonly contains scattered lenses of hornblende-biotite amphibolite a few metres thick. This well-layered gneiss grades upward into wispy and streaked gneiss and finally into strongly foliated but generally unlayered gray biotite-quartz-plagioclase gneiss that lacks both hornblende and/or microcline characteristic minerals of other foliated gray gneisses in the section.

The calc-silicate unit underlies the biotite-quartz-plagioclase gneiss and contains several distinctive but apparently lenticular rock units. At the base of the calc-silicate are lenses of dark-green to black hornblende-biotite-diopside amphibolite and, locally, rusty weathering graphitic thin-bedded calcareous quartzite. These rocks are overlain by well-layered green, white, and red diopside-microcline-garnet gneiss. Essentially monomineralic garnet layers locally attain a thickness of 1 m. Pods of coarsely crystalline calcite-diopside-chondrodite marble occur at various levels in the calc-silicate unit.

The calc-silicate unit is underlain by well-layered black and white hornblende-biotite-plagioclase gneiss. Hornblende-rich and plagioclase-rich layers vary in thickness from a few centimetres to a few metres.

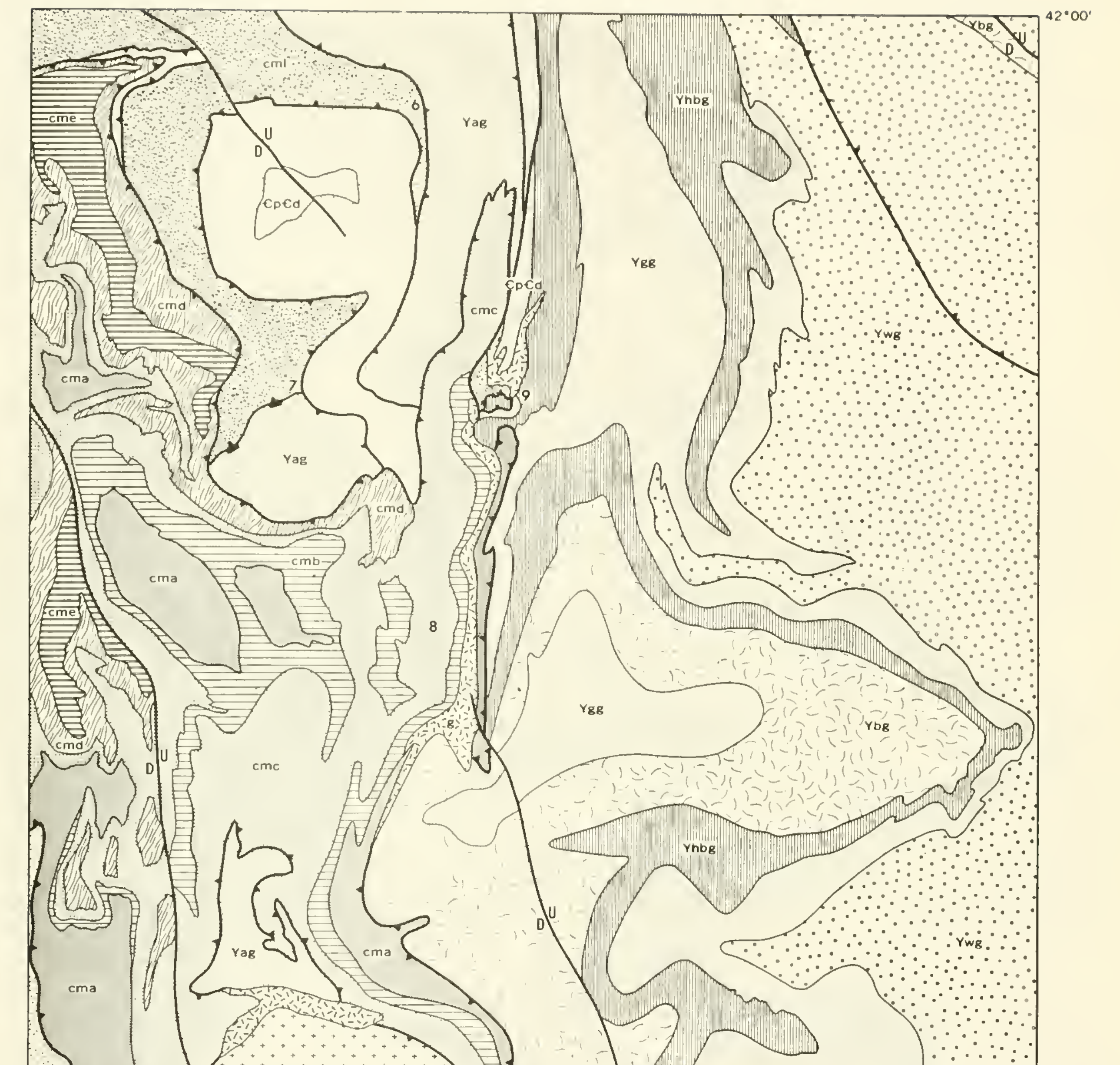
The hornblende-biotite-plagioclase gneiss is in sharp contact with strongly foliated but generally unlayered coarse grained biotite-ferrohastingsite granite gneiss. Biotite is the dominant mafic mineral and forms discontinuous laminations a few millimetres thick in a strongly foliated groundmass of about equal amounts of

quartz, plagioclase, and microcline. Ferrohastingsite is present in minor amounts and distinguishes this granite gneiss from the otherwise similar biotite-muscovite granite gneiss in the southern part of the South Sandisfield quadrangle and the adjacent Norfolk quadrangle. The biotite-ferrohastingsite granite gneiss is similar in mineralogy and texture to the Tyringham Gneiss of Emerson (1899).

The biotite-ferrohastingsite granite gneiss is in sharp, locally chilled, contact with rusty-weathering biotite-muscovite-sillimanite schist and gneiss which forms an extensive unit in the eastern part of the area. These rocks are readily distinguished by their abundance of muscovite, an uncommon mineral in the other Precambrian rocks, their rusty-brown or orange weathering rind, and by the distinctive quartzite beds in the schist that weather to a gray vitreous lacework pitted surface. All these features are characteristic of the Washington Gneiss of Emerson (1899); the only difference between the two is the absence of blue quartz found in the Washington Gneiss at lower metamorphic grades to the north. Along the eastern side of the South Sandisfield quadrangle is a unit of biotite-quartz-plagioclase hornblende gneiss in or below the rusty schist and gneiss of the Washington.

The Precambrian rocks below the Beartown Mountain thrust (fig. 2) in the south-central part of the South Sandisfield quadrangle and the eastern part of the Norfolk quadrangle are very similar to those of the calc-silicate-bearing sequence described above, but the calc-silicate marker unit is missing. For this reason, the correlation of units above and below the Beartown Mountain thrust cannot be considered above reproach, but there are enough similarities to merit at least tentative correlation. The rusty-weathering biotite-muscovite-sillimanite schist and gneiss of the Washington Gneiss and the biotite-quartz-plagioclase gneiss associated with it appear unchanged below the Beartown Mountain thrust. Below the Washington Gneiss is a unit of medium- to coarse-grained, strongly foliated but generally unlayered biotite granite gneiss. This granite gneiss is very similar in appearance and at the same stratigraphic position as the biotite-ferrohastingsite granite gneiss, but there are large tracts in which it contains minor amounts of muscovite and no amphibole. Other parts of the granite gneiss contain ferrohastingsite and no muscovite. These two types of granite gneiss could not be separated in the field. Well-layered, black and white hornblende-biotite-quartz-plagioclase gneiss underlies the biotite granite gneiss. Biotite-quartz-plagioclase gneiss, somewhat better layered than the similar rock above the Beartown Mountain thrust, underlies the black and white hornblende-biotite gneiss. If the sequences of rocks above and below the Beartown Mountain thrust were exactly the same, the calc-silicate unit should appear between the well-layered hornblende-biotite gneiss and the biotite-quartz-plagioclase gneiss. Calc-silicate pods and patches of hornblende-biotite-diopside rock do occur in the hornblende-biotite gneiss, and these may represent the well-developed calcareous horizon above the Beartown Mountain thrust.

Figure 2. Generalized geologic map of the South Sandisfield quadrangle (above) and Norfolk quadrangle (right).


42°00′
41°52′30″
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73°07′30″
cml
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g
U
D
6
7
8
9
0
1 MILE
0
1 KILOMETRE

EXPLANATION

LOWER CAMBRIAN TO MIDDLE ORDOVICIAN(?) OR YOUNGER

ORDOVICIAN (?)

g

Granite

ORDOVICIAN

OЄru

Stockbridge and Walloomsac Formations, undivided

UPPER PRECAMBRIAN(?) AND LOWER CAMBRIAN

PRECAMBRIAN(?) AND CAMBRIAN

ЄpЄd

ЄpЄdg

Dalton Formation

ЄpЄd - feldspathic granulite and schist

ЄpЄdg - flaggy-bedded quartzite

CANAAN MOUNTAIN SCHIST OF ROGERS AND OTHERS (1959)

cme

cmd

cmc

cmb

cma

Upper Canaan Mountain slice

cma - feldspathic micaceous granulite

cmb - mafic volcanic rock

cmc - mucovite-spangled gneiss

cmd - garnet, sillimanite, magnetite schist

cme - well bedded feldspathic granulite

cml

Lower Canaan Mountain slice

Feldspathic, micaceous granulite and garnet, sillimanite schist

ЄpЄw

Waramaug Formation of Gates (1952, 1965)

Feldspathic, micaceous granulite with abundant amphibolite lenses, minor garnet, sillimanite schist

PRECAMBRIAN Y

PRECAMBRIAN

Yag

Haystack Mountain slice

Hornblende-biotite amphibolitic gneiss

Ybg

Ycs

Yhbg

Ygg

Ywg

Ybg

Gneisses of the massif

Ybg - biotite-quartz-plagioclase gneiss

Ycs - calc-silicate gneiss

Yhbg - hornblende-biotite gneiss

Ygg - granite gneiss

Ywg - Washington Gneiss and biotite-quartz-plagioclase gneiss (Ybg)

Right-side up

Overturned

Thrust fault

D

U

Normal fault

1

Stop locality

Rocks of the Haystack Mountain slice (fig. 1) are black and white well-layered hornblende-biotite-plagioclase gneiss containing mappable layers of amphibolite several metres to several tens of metres thick. In general, this gneiss is more amphibolitic than other black and white gneisses in the area. The Haystack Mountain slice appears to be completely separated from the other Precambrian gneisses by faults and by a thin sliver of the Dalton Formation. Because it contains essentially one mappable unit, there is little information to suggest where it fits in the Precambrian sequence. It could be equivalent to the other black and white gneiss or it could be far travelled and totally unrelated to any of the adjacent gneisses.

Stratigraphy of the Dalton Formation and the Canaan Mountain Schist of Rodgers and others (1959)

The Dalton Formation, which normally lies conformably beneath the Cheshire Quartzite (Lower Cambrian) and forms the lowest unit of the autochthonous platform sequence, occurs both in separate thrust slices structurally above the platform sequence and as unconformable cover rocks to the Precambrian gneiss in other thrust slices in the area of this trip. On Higley Hill, the basal 3 m of the Dalton consists of coarse-grained granular schist composed of abundant muscovite, two feldspars, quartz, biotite, and tourmaline. This rock, which may be a metamorphosed saprolite, grades upward into tan-weathering, medium-grained, sugary-textured feldspathic quartzite containing beds a few centimetres to a few metres thick. Muscovite-tourmaline-rich partings separate the felspathic quartzite beds and give the rock a characteristic flaggy appearance. The flaggy-bedded quartzite is the lowermost unit of the Dalton exposed at Campbell Falls (stop 4), and it is overlain by coarse-grained biotite-muscovite-sillimanite schist. The schist contains mappable layers of quartzite. Locally, near the base of the schist, is a strongly foliated feldspathic, micaceous granulite that contains distinctive almond-shaped white knots of quartz, muscovite, and sillimanite. These metamorphic segregations give the rocks a pseudoconglomeratic appearance.

Two thrust slices on Canaan Mountain (fig. 1) contain micaceous granulites remarkably similar in appearance and mineralogy to the micaceous granulites of the Dalton Formation. Rocks associated with the micaceous granulites on Canaan Mountain, however, are significantly different from rocks of the Dalton. The lower thrust slice on Canaan Mountain contains tan to gray, strongly foliated, poorly bedded micaceous granulite at the base. This granulite unit contains scattered lenses of amphibolite as much as 15 m thick. Coarse-grained, porphyroblastic biotite-muscovite-garnet-sillimanite-staurolite schist overlies the lower micaceous feldspathic granulite.

The upper thrust slice on Canaan Mountain contains a more varied stratigraphy than either the lower Canaan Mountain slice or the Dalton Formation. The lowermost unit is poorly bedded, strongly foliated, tan-weathering feldspathic and micaceous granulite that contains almond-shaped quartz-muscovite-sillimanite knots. This rock is virtually indistinguishable from parts of the Dalton, but it is overlain by black to dark-gray biotite-hornblende-quartz-plagioclase gneiss that contains white porphyroblasts of albite or sodic oligoclase. This unit also contains some massive amphibolite beds a few metres thick suggesting it originated as a mixture of mafic volcanic flows and pyroclastic deposits. Above the volcanic rocks is a distinctive steel-gray to black coarse-grained muscovite-biotite-quartz-microcline gneiss. This micaceous gneiss is strongly foliated and contains aggregates of coarse-grained muscovite that look like fish scales smeared out in the biotite-rich folia. Above the micaceous gneiss is medium- to coarse-grained garnet-sillimanite-staurolite schist that contains abundant quartz stringers and finely disseminated magnetite. The abundance of magnetite suggests that this gray to tan schist was once a red hematite-rich pelitic sediment. Red plates of hematite are, in fact, trapped in some of the garnets. Above the magnetite-rich schist is a unit of well-bedded tan feldspathic quartzite. This quartzite contains biotite, muscovite, quartz, microcline, and plagioclase much like the Dalton, but the bedding is less distinct, the slabby parting parallel to bedding is missing, and this rock contains considerably less quartz than does the typical feldspathic quartzite in the Dalton.

In general, the feldspathic micaceous granulite in the Canaan Mountain slices closely resembles the granulite in the Dalton, including the muscovite-quartz-sillimanite segregations common to both. In contrast to those of the Dalton, the schists in the Canaan Mountain slices are more aluminous, more iron-rich, and contain abundant thin stringers of quartz. The schist and granulite of the Dalton contain quartz and feldspar segregations, but these segregations tend to be contorted knots studded with black tourmaline. The relatively quartz-rich flaggy-bedded quartzites of the Dalton are missing from the Canaan Mountain sequence, which contains more massive, more feldspathic arenaceous rocks. The amphibolite lenses and pyroclastic rock are missing in the Dalton. The Dalton north of the trip area contains quartz pebble to cobble conglomerate beds, whereas the arenaceous rock in the Canaan Mountain slices are medium- to coarse-grained sand containing no pebble layers. Compared with those of the Dalton, therefore, the rocks of the Canaan Mountain sequence appear to be deeper water deposits laid down east of the carbonate platform within range of volcanic flows and pyroclastic deposits possibly originating from fissures and volcanoes in a marginal basin-island arc setting.

Fold-Thrust Structures in the Precambrian Rocks

In the northwestern part of the South Sandisfield quadrangle, the calc-silicate unit outlines some strongly overturned to recumbent folds which are broken and truncated by generally east-dipping thrust faults. The thrust faults cut the calc-silicate-bearing sequence into a series of imbricate thrust slices so that the lower slices are successively overlapped to the north and east by the next higher slices, much like shingles on a roof. Within each thrust slice, the relatively older rocks tend to be in contact with the thrust fault at the base of the slice and form the cores of westward- or southward-closing anticlines. The relatively younger rocks in the calc-silicate sequence either form the troughs of major overturned synclines or they appear as synclinal digitations on the overturned sheared-out limbs of the recumbent anticlines adjacent to the thrust faults. Within a few tens of metres of the thrust faults is a pronounced cataclastic foliation, described in detail by Ratcliffe and Harwood (1975), that lies parallel to both the axial surfaces of the overturned folds and to the thrust faults. This penetrative cataclastic foliation clearly relates the recumbent folding to the thrusting. The general fold-thrust style of deformation is shown by the composite cross section in figure 3.

Near Higley Hill (fig. 2), the calc-silicate unit outlines the hinge of a major recumbent syncline that contains the Dalton Formation in its trough. The normal limb of this syncline follows the trace of the calc-silicate unit westward to Benton Hill (fig. 2), where successively lower units in the sequence are exposed. On the west slope of Benton Hill (stop 1), biotite-ferrohastingsite granitic gneiss rests on schist and marble of the Walloomsac Formation (Middle Ordovician or younger). The granitic gneiss is in the core of a recumbent anticline in which the lower limb is sheared out along the Benton Hill thrust.

The next higher thrust slice (slice C, fig. 1) truncates the Dalton Formation in the core of the recumbent syncline northwest of Higley Hill and also truncates the Benton Hill thrust west of the town of Southfield. A tectonic sliver of Walloomsac marble is caught between rocks of the Benton Hill thrust slice and intensely sheared biotite-ferrohastingsite at the base of the Harmon Pond slice (slice C, fig. 1) about a kilometre south of the town of Southfield. Within the Harmon Pond slice, the Precambrian rocks are in normal stratigraphic succession; and west of Southfield, the biotite-quartz-plagioclase gneiss forms the trough of a tightly appressed recumbent syncline that flairs out to the southeast into a series of strongly overturned folds. The map pattern east of Harmon Pond (fig. 2) is complex because northwest-trending, westward-overturned folds that formed during the thrusting are superposed on an earlier, probably Precambrian, fold that trends eastward from Harmon Pond.

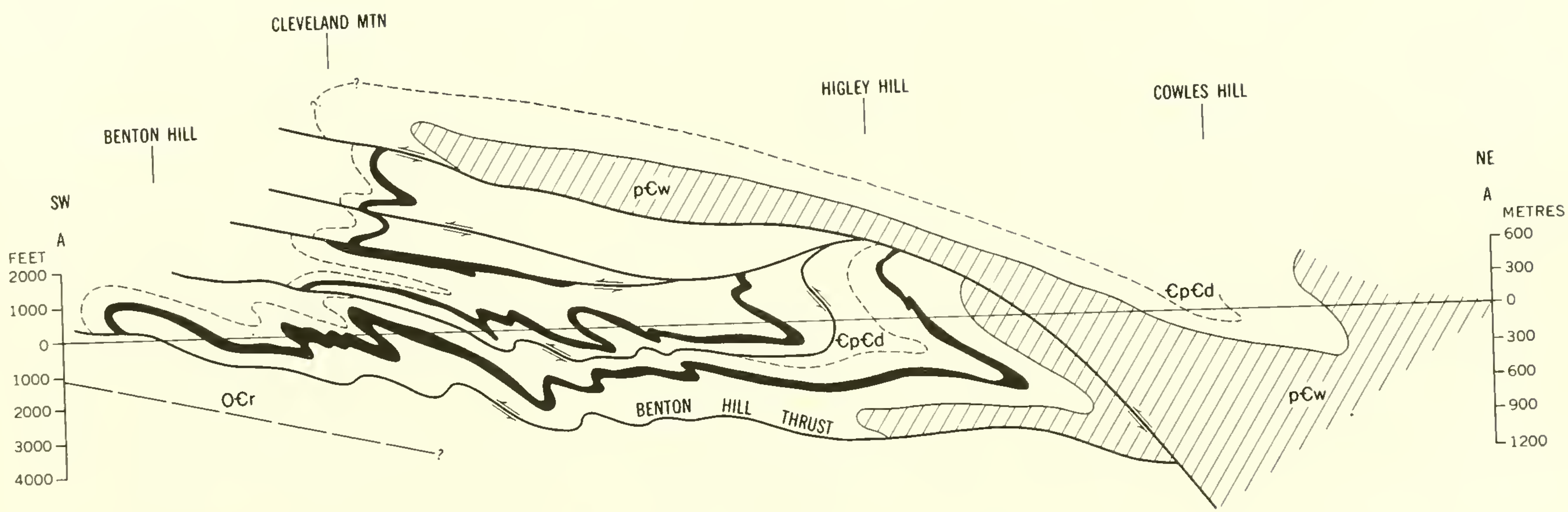

Figure 3. Composite cross section along A-A' (fig. 2) showing fold-thrust structures in the South Sandisfield quadrangle; calc-silicate unit (black); Washington Gneiss (p€w, ruled); dotted line is unconformity below Dalton Formation (€p€d); undivided carbonate platform rocks (O€r).

Between Cook's Ledge and Woodruff Mountain (fig. 2), the calc-silicate unit in thrust slices D and E outlines partial or complete synclinal digitations on the overturned limb of a major recumbent anticline cored by the Washington Gneiss. The form of these synclinal digitations and their relationship to the thrust faults are best shown in the composite cross section (fig. 3).

The total amount of shortening involved in the thrusting and recumbent folding is difficult to determine accurately. An order-of-magnitude estimate of the minimum shortening, however, can be obtained if we assume that the patches of Dalton Formation on Cleveland Mountain, Higley Hill, and Cowles Hill were once part of a continuous blanket of cover rocks resting unconformably above the Precambrian gneisses. The present horizontal distance between the patches of Dalton on Cleveland Mountain and Cowles Hill is about 8 km. The original distance between these two patches of Dalton, measured around the dotted line on figure 3, is about 28 km or a shortening of about 20 km. This amount of shortening does not include the movement on the fault at the base of the Benton Hill slice (slice B) because the eastward extent of carbonate platform rocks beneath that slice is not known. A minimum shortening of 20 km, however, compares very well with the minimum transport distance of about 21 km on the Beartown Mountain slice to the north (Ratcliffe and Harwood, 1975).

Fold-thrust Structures in the Canaan Mountain Rocks

There is not enough lithologic variation in the lower Canaan Mountain thrust slice to determine the geometry of any large-scale fold-thrust structures, if any exist. The repetition and attitude of distinctive lithologic units in the upper slice, however, indicates the presence of large-scale recumbent folds subsequently refolded about northwest- and northeast-trending axial surfaces. A pronounced schistosity lies parallel to the axial surfaces of the recumbent folds and deforms an earlier, less distinct foliation produced by fine-grained muscovite and biotite. Garnet appears to have formed initially during the metamorphism that produced the early foliation and subsequently grew inclusion-free rims during sillimanite-grade metamorphism following the thrusting. Sillimanite and staurolite lie in the fold-thrust fabric and formed by the breakdown of biotite and, to a lesser extent, muscovite. The dominant schistosity becomes a blastomylonitic foliation in the thrust zones.

There are two major recumbent folds in the upper Canaan Mountain thrust slice. The upper fold, shown as the Crissey Pond fold on figure 1, is a northwest-closing anticline with feldspathic, micaceous granulite containing quartz-sillimanite knots much like those in part of the Dalton Formation in its core. The lower fold, the Camp Pond fold on figure 1, is a southeast-closing syncline containing well-bedded, feldspathic and micaceous quartzite in its trough. These major folds are domed upward around the Tobey Pond window (fig. 1), deformed by north-trending folds related to the upthrusting of Precambrian rocks east of Dennis Hill and Spaulding Pond, and are clearly truncated by the

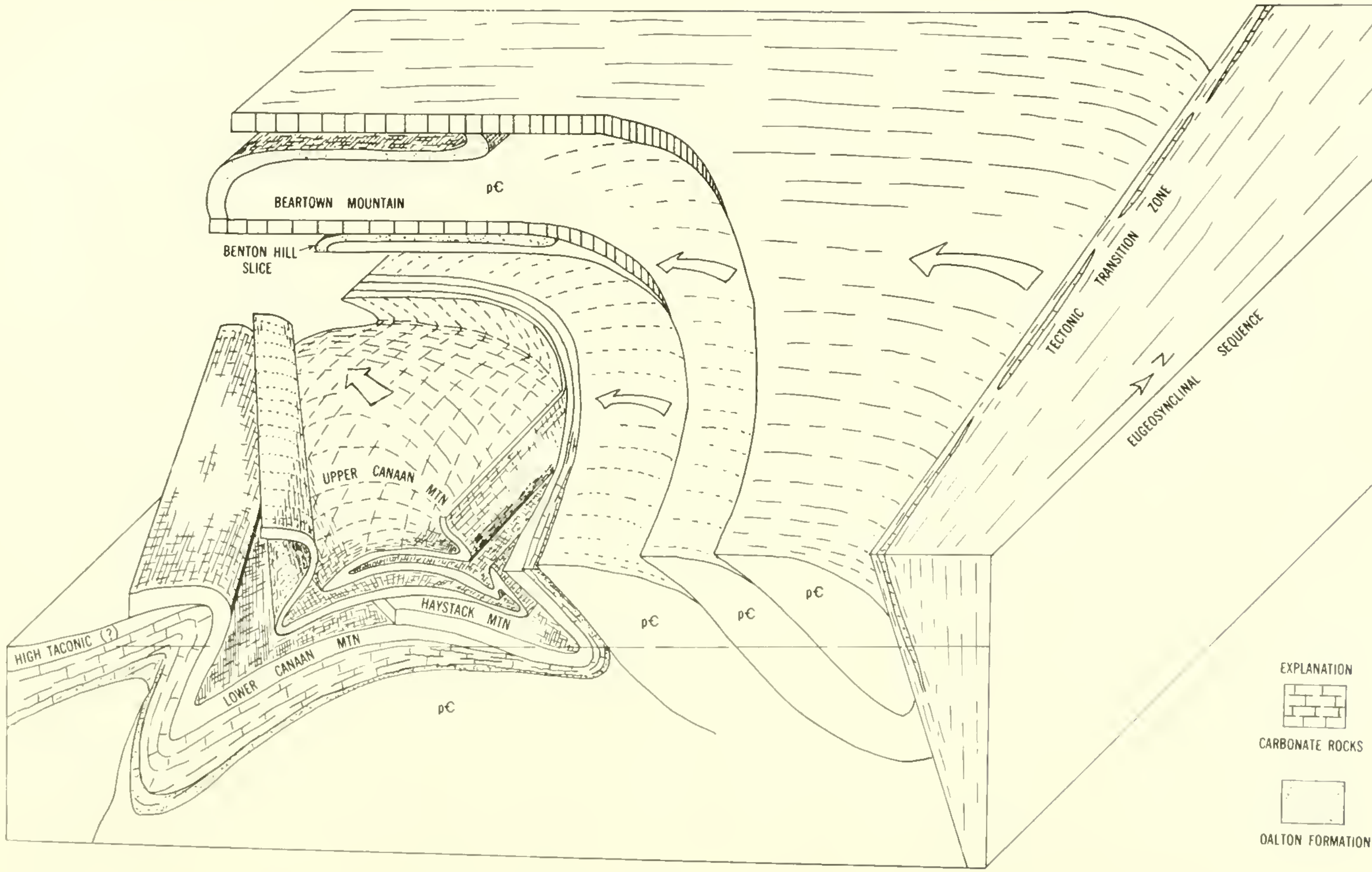

Figure 4. Block diagram showing stacking order of thrust slices and major folds formed during thrusting of Canaan Mountain rocks and Precambrian rocks of the Berkshire massif. Dashed line on front face of diagram approximately locates present erosion level; large arrows show possible directions of tectonic transport.

parautochthonous Waramaug Formation of Gates (1952, 1965) south of the North Pond windows (fig. 1). The period of fold-thrust tectonism, therefore, appears to have progressed through a series of deformational events, all preceding the late northwest- and northeast trending folds that warped the thrust slices.

Speculations on the fold-thrust tectonic events

The structural relations at Stop 9 and those to the west on Canaan Mountain indicate that the Canaan Mountain Schist is in fault contact with the carbonate sequence. Major fold-thrust structures in the upper Canaan Mountain thrust slice are truncated to the south by the Waramaug Formation. The lithologic similarity between units in the Canaan Mountain Schist and parts of the Dalton, Waramaug, Hoosac, and possibly the Everett and Rowe suggest that the Canaan Mountain rocks were originally part of the eugeosynclinal sequence deposited east of the Berkshire massif. I conclude that the Canaan Mountain rocks are allochthonous and probably part of the Taconic allochthon.

On Canaan Mountain, the recumbent folds related to the fold-thrust deformation are overturned to the northwest, whereas similar recumbent folds outlined by the calc-silicate unit in the Precambrian rocks are overturned to the west and southwest. Although no slip line directions have been determined for any of these folds, the direction of overturning and the significantly different stacking directions to the north and south suggest that the Canaan Mountain slices were transported from the southeast and that the Precambrian slices moved from the east.

The eastern part of the Canaan Mountain fold-thrust structures are refolded into a north-trending, westward-overturned synform against the upthrust Precambrian block to the east. The Canaan Mountain slices and the Haystack Mountain slice had to be emplaced before upthrusting of the Precambrian rocks. The geometry of the refolded Canaan Mountain slices and their relationship to the Precambrian rocks are shown in figure 4.

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Road Log for Trip B2 (Sat.) and C2 (Sun.)

Mileage	
0	Trip starts in the parking lot of the Monument Mountain Regional High School and proceeds south through the Great Barrington, Ashley Falls, South Sandisfield, and Norfolk quadrangles.
	Leave the parking lot and turn left on route 7.
0.2	Turn left on Monument Valley Road.
4.9	Cross route 23 - continue south on Lake Buel Road.
7.5	Turn right on Great Barrington Road.
9.3	Junction Mill River - Sheffield Road on right - continue south on Great Barrington Road to Mill River.
11.0	Town of Mill River. Turn left and follow Mill River Road uphill.
11.8	Four-way junction. Keep to the right and follow Hadsell Street.
12.6	Bridge-road junction. Cross bridge and turn right on Canaan-Southfield Road.
14.1	Turn left on Cross Road (dirt) and proceed uphill.
14.7	STOP 1. West slope of Benton Hill (Ashley Falls quad.). These outcrops have been graciously loaned to me by N. M. Ratcliffe and H. R. Burger who mapped the Ashley Falls quadrangle. Outcrops beside the road are tan-weathering micaceous marble, the lowermost unit of the Walloomsac Formation. Follow trail uphill to ledges and blocky talus of black biotite-splotched schist containing abundant quartz stringers, a few calcareous feldspathic quartzite beds, and scattered quartz-feldspar pods. This is the black slate member of the Walloomsac at sillimanite grade. Note the early asymmetrical folds overturned to the west; pronounced axial-plane schistosity strikes north and dips gently east. Later crenulation cleavage strikes about N.15°E. and is nearly vertical. Carefully continue uphill over scattered outcrops of the pelitic unit of the Walloomsac across a small swale to the ledge of biotite-ferrohastingsite granite gneiss. The granite gneiss is very similar to and probably akin to the Tyringham Gneiss dated at 1.06 b.y. (Ratcliffe and Zartman, 1971). The rock is normally a strongly foliated, pinkish-gray, coarse-grained, porphyritic granite gneiss. Here, a few metres above the Benton Hill thrust, the granite gneiss is strongly crushed and crossed by gently east-dipping dark gray to black zones

of blastomylonite and an anastomosing cataclastic foliation described in detail by Ratcliffe and Harwood (1975). Remnants of the Precambrian foliation are preserved in the blastomylonite zones. Proceed north along the base of the granite gneiss ledges to the point where the contact between the Walloomsac marble and the granitic gneiss is exposed in a tight isoclinal upright fold. Note the idocrase at the contact and the fine grain size of the granitic gneiss. There is a strongly overturned fold outlined by pegmatic material in the granite gneiss in the ledges above the contact.

Retrace the route back to the cars. Turn the cars around and proceed west on Cross Road.

15.4 Turn right on Canaan-Southfield Road.

16.8 Junction of milestone 12.6 - continue east on Canaan-Southfield Road. DO NOT CROSS BRIDGE.

17.7 Four-way intersection--red house as landmark in the northwest quadrant. Continue east on Canaan-Southfield Road.

18.0 Turn right on Foley Hill Road. Sign to Xmas Tree Farm. Proceed uphill on dirt road.

19.2 Gate to YMCA camp on left; enter and park. There is a short hike to STOP 2.

There are scattered pavement outcrops of calc-silicate rock where the cars are parked; better exposures follow. Walk east (downhill) to the north end of Harmon Pond; cross the outlet stream and walk north about 100 m to outcrops of hornblende-biotite gneiss. The thrust fault is in the brook. Note gently east-dipping cataclastic foliation and the small scale westward-overturned folds sheared out along their overturned limbs. Within the cataclastic foliation are several detached fishhook-shaped folds outlined by the Precambrian foliation. Retrace the path to Harmon Pond and continue walking east along the water's edge. A variety of calc-silicate rocks are well exposed on the low hills east of the pond. They include white microcline gneiss spotted with coarse diopside and sphene, massive layers of garnet, diopside-garnet-quartz-microcline gneiss, and diopside-calcite-quartz marble pods.

Retrace the trail back to the cars. Turn the cars around and proceed down Foley Hill Road to the junction with Canaan-Southfield Road.

20.4 Turn right (east) on Canaan-Southfield Road.

21.0 Dirt road spurs off to the right - continue on paved road.

21.3 Junction with route 272 in the town of Southfield. Turn right and park the cars on the side of the road.

STOP 3 - CAUTION - The talus blocks are treacherous and parts of the overhanging ledges are metastable at best. Outcrops on the east side of the road are black and white hornblende-biotite gneiss in fault contact with the calc-silicate rock, which is exposed in a few places along the base of the ledge. Prominent features to note here are the recumbent folds in the black and white gneiss, the thin zones of cataclasis and granulation, minor fault surfaces with quartz stringers and slickensides and the late north-trending upright folds that deform the fold-thrust structures. At the eastern end of the ledge, layering in the gneiss outlines the hinge of a major recumbent fold that has been sheared out along the thrust fault.

Return to the cars and continue south on route 272.

24.9 Turn right (west) on dirt road to Campbell Falls. Massachusetts-Connecticut State line is landmark.

25.4 STOP 4 - LUNCH - Campbell Falls. Thin-bedded feldspathic quartzite containing muscovite-rich partings holds up the falls and is flanked to the west by coarse biotite-muscovite-sillimanite schist. Both rock types are in the Dalton Formation. The feldspathic quartzite beds outline the hinge of a late postthrusting anticline in which the axial surface dips steeply east and the axis plunges about 20°S. At the base of the falls the river has exposed a bedding surface on the hinge of the fold that contains abundant tourmaline needles. These tourmaline needles define an early lineation, possibly formed during thrusting, that was refolded into a helical pattern by the late anticline. It is left as an exercise for the interested student to gather and plot the tourmaline lineation data. Also at the base of the falls is an exposure of micaceous granulite which contains almond-shaped white quartz-sillimanite-muscovite knots; the pseudo-pebbles discussed by Emerson (1899, p. 67).

Return to the cars and continue west on Campbell Falls Road.

26.8 Junction with Canaan Valley Road - cross bridge.

26.9 Turn left (south) on Canaan Valley Road.

27.4 Junction Cross Road - continue south.

27.6 Turn left (east) on Toby Hill Road.

27.7 Junction - take left fork - uphill.

28.7 Pass Yale Farm on the right.

29.3 STOP 5. We are about a quarter of a mile south of Campbell Falls, still in the Dalton Formation and very close to the contact with the underlying Precambrian gneiss. The contact is not exposed here, but it is inferred to be a thrust fault from regional relationships. Micaceous granulite with abundant quartz-sillimanite knots contains a thin sill of two-mica granite that is essentially parallel to the early foliation in the Dalton. The granite sill and the early foliation are folded into a westward-overturned syncline that has the fold-thrust fabric as its axial surface. The quartz-sillimanite knots are parallel to the early foliation on the limbs of the syncline and become progressively deformed into sigmoids and are eventually flattened in the plane of the fold-thrust fabric near the axial surface of the syncline. Microscopically, the knots are aggregates of strained quartz surrounded by wisps of sillimanite replacing biotite. Large muscovite flakes at the edges of the knots are deformed and crossed by streaks of sillimanite. The groundmass is largel microcline, quartz, and biotite. The textures indicate that the knots, the sill, and the early foliation were present in the rock before the fold-thrust event. The age of the sillimanite with respect to the thrusting is not clearcut; the sillimanite appears to have formed from the breakdown of biotite either during or after thrusting.

Return to the cars and continue east on Toby Hill Road.

29.7 Turn right (south) on route 272.

33.9 Junction of routes 272 and 44 in Norfolk, Connecticut - continue south.

34.2 Junction of routes 44 and 72 - proceed south on route 72.

34.4 Turn right (west) on Mountain Road in Norfolk village.

34.5 Turn right onto grounds of Yale Music School and park by the large concert hall.

STOP 6 - Tan to gray strongly foliated feldspathic, micaceous granulite at the base of the lower thrust slice on Canaan Mountain. The granulite grades upward into garnet-sillimanite schist. The purpose of this stop is to compare the arenaceous and pelitic rocks of the Canaan Mountain sequence with those of the Dalton Formation seen at stops 4 and 5.

Return to Mountain Road and turn right (west).

34.9 Turn right on Westside Road.

36.7 STOP 7 - Tan, feldspathic, and micaceous granulite in the lower Canaan Mountain thrust slice. Compare this pseudoconglomeratic granulite with that of the Dalton seen at stop 5.

37.8 Turn right (south) on route 72.

38.1 Enter Dennis Hill State Park and proceed to the pavilion on top of Dennis Hill.

STOP 8 - Steel-gray, muscovite-rich, biotite-quartz-microcline gneiss that forms a widespread unit in the upper Canaan Mountain thrust slice. In the service road east of the pavilion, the gneiss is intruded by a mafic dike, one of several in this zone of faulting near the contact with the Precambrian rocks, composed of diopside-garnet-hornblende-biotite-sodic plagioclase-quartz-magnetite and possibly ilmenite. Garnet is completely rimmed by quartz and plagioclase, and diopside and plagioclase are intergrown in a symplectite texture. The present mineralogy is peculiar for a mafic intrusive; the textures suggest the rock was originally composed largely of garnet and omphacite and thus was an eclogite. The jadeitic component of the omphacite is now present as sodic plagioclase intergrown with the diopside component.

Return to route 72 and proceed north toward Norfolk.

38.7 Turn right toward Lake Winchester and Winsted.

39.2 STOP 9 - Park on the side of the road and hike about 0.5 mile north to the steep hills at the south end of Spaulding Pond. The cars are parked on a narrow body of granite which intrudes rocks of the upper Canaan Mountain thrust slice. On the hike to stop 9, we will cross over hornblende-plagioclase gneiss also intruded by several small dikes of granite and see the Dalton Formation, about 1 m of relatively clean quartzite that may represent the Cheshire Quartzite, and about 1.5 m of coarse-grained white dolostone that is part of unit a of the Stockbridge Formation. These rocks of the carbonate platform sequence are intensely folded about north-trending axes and plunge northward beneath black and white hornblende-biotite gneiss. The Stockbridge is in obvious fault contact with strongly foliated feldspathic, micaceous granulite like that in the lower Canaan Mountain thrust slice seen at stop 6. Fine- to medium-grained two-mica granite forms north-trending dikes in the fault zones. The map pattern indicates that the carbonate platform sequence is caught in the heel of a north-trending, north-plunging, westward-overturned syncline capped by the trailing edge of the lower Canaan Mountain thrust slice. Vertical faulting cut the heel of

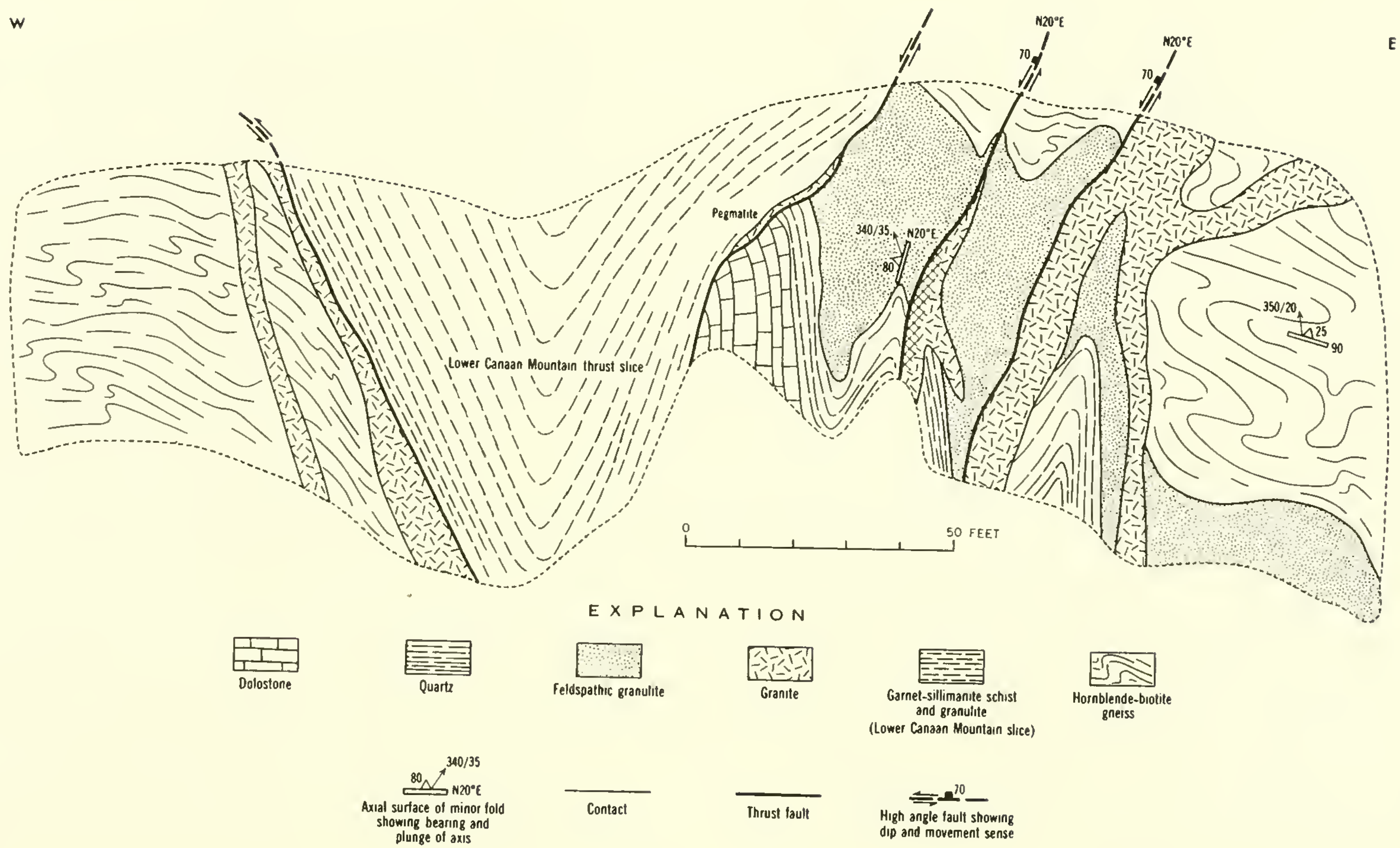

Figure 5. Sketch of near vertical ledge at stop 9 (south of Spaulding Pond; Norfolk quadrangle) showing fault contact between lower Canaan Mountain thrust slice and dolostone of the Stockbridge Formation. "Quartz" in the explanation should read quartzite.

the syncline after the thrusting and granite intruded along the vertical faults. The pattern of map units is shown in figure 5.

Return to the cars - retrace the route to highway 72 - turn north (right) to Norfolk.

The best way to get to Great Barrington is route 44 from Norfolk to Canaan and then north on route 7

Proposed Silurian-Devonian correlations east of the Berkshire massif in western Massachusetts and Connecticut

by

Norman L. Hatch, Jr. and Rolfe S. Stanley

In USGS Bulletin 1380 (Hatch and Stanley, 1973), we have proposed that most of the Straits Schist and parts of other units in the belt of rocks immediately west of the Triassic basin in western Connecticut are Silurian and Devonian in age and correlative with the Goshen Formation of Massachusetts (Hatch, 1967). These correlations are based on lithic similarity and on the stratigraphic sequence in which the rocks occur. Because many workers in western Connecticut have placed the Straits Schist in the middle of a presumed Cambrian and Ordovician section, our suggested placement of the Straits unconformably above all of the other crystalline stratified units in western Connecticut requires a major revision of the structure.

Figure 1 shows the route of the trip and the stops on a geologic base which shows our interpretation simplified from plate 1 of Bulletin 1380. From figure 1 it can be seen that the Goshen Formation in Massachusetts, along with the underlying Russell Mountain Formation (Hatch, Stanley, and Clark, 1970), pass under the Triassic cover west of Westfield, Mass. The rocks therein shown as Russell Mountain and Goshen-equivalent Straits around the Granville and Granby domes are separated from presumably equivalent rocks to the north by pre-Silurian rocks. They are also separated from their presumed stratigraphic correlatives to the south in the Collinsville, Conn. area by Cambrian and Ordovician rocks. This intervening area has been mapped by R. W. Schnabel (Schnabel and Eric, 1965; Schnabel, in press, 1974, 1973) who mapped the rocks as pre-Silurian Straits Schist. Although Schnabel does not subscribe to the reinterpretation presented here, he has generously guided us through his area on many occasions.

The Straits Schist forms a continuous mappable belt south from the Collinsville and Bristol domes of the Collinsville quadrangle to Long Island Sound. Thus if our stratigraphic-structural interpretation for the Collinsville area is accepted, it would also apply south to the Sound.

Critical to the development of our interpretation of the Connecticut rocks has been the Silurian Russell Mountain Formation (Hatch, Stanley, and Clark, 1970). This thin but distinctive unit of calc-silicate rock and quartzite is presumably correlative with the Shaw Mountain Formation of Vermont (Doll and others, 1961). The Russell Mountain Formation first appears in Massachusetts at Blandford Village and is mapped discontinuously to the south. Our proposed correlation of most of the Straits with the Goshen of Massachusetts is greatly fortified, we believe, by the presence in western Connecticut of a narrow belt of calc-silicate rock and quartzite structurally and apparently stratigraphically thought to be correlative with the Silurian and Devonian Goshen Formation and rocks that we correlate with the Hawley (Hatch, 1967) and Cobble Mountain (Hatch and Stanley, 1973) Formations.

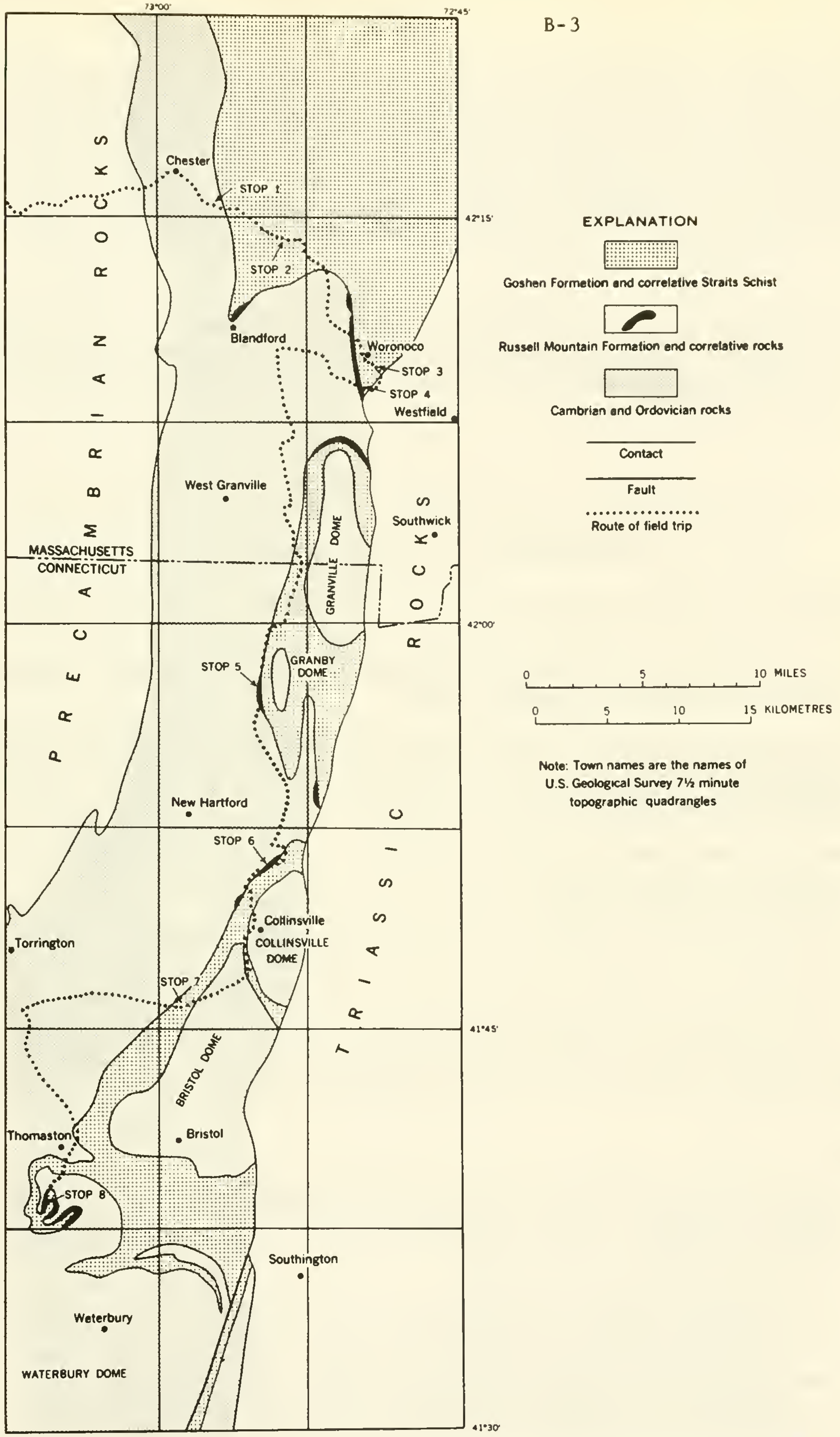

Figure 1. Sketch map of part of western Massachusetts and Connecticut showing distribution of proposed Silurian and Silurian-Devonian rocks and field trip stops. Simplified from plate 1 of Hatch and Stanley (1973).

The purpose of this trip is to show in the field some of the evidence upon which we based our correlations and interpretation. The relations are complex and some of the lithic distinctions are subtle. We will attempt to show in eight stops a story that has evolved from a combined total of over 20 years of work in these rocks. Bear with us, look at the rocks, listen to our story, and make your own decisions. The interested student of these problems should read Bulletin 1380 where many of the points are discussed in much greater detail.

Finally, but not least important, we have taken great liberties with the geology mapped by many of our colleagues in western Connecticut. We acknowledge the excellence of their work and hasten to point out the very obvious fact that without their long and arduous efforts no reinterpretation such as this could be possible.

Instructions to all participants

Assemble at Monument Mountain Regional High School, Great Barrington, at 7:30 A.M.

Lunches will <u>not</u> be provided. Bring your own lunch, please. No provision can be made for stopping at grocery stores.

Leave Great Barrington at Route 7 proceeding north to Stockbridge. Take Route 102 east to Lee where you will join Route 20. Proceed east on Route 20 through East Lee and Chester.

Stop 1 is approximately 1.7 mi. southeast of the center of Chester Village.

<u>Note other warning signs</u>

(a) Round Hill Road southeast of Chester is 0.95 mi. northwest of Stop 1.

(b) Blair Brook crosses Route 20 about 0.6 mi. northwest of Stop 1.

Stop 1 is marked by a small steel bridge spanning the West Branch of the Westfield River. Rest Area is located directly northwest of bridge. Park to the north of road.

<u>Mileage</u>

0.0 <u>STOP 1</u>. 150 m east on Route 20 of narrow bridge over West Branch of Westfield River. Park in Rest Area on left (north) side of road. Chester quadrangle (Hatch, Norton, and Clark, 1970). Outcrop is to east along south side of road.

The outcrop here consists of rocks mapped as carbonaceous schist units of the Hawley Formation by Hatch, Norton, and Clark (1970). the west end of the outcrop consists of brown, sandy nongraphitic medium- to fine-grained quartz-muscovite-biotite-garnet schist interbedded with rusty-, splintery-weathering, gray, graphitic,

sulfidic, fine-grained, slabby quartz-biotite-muscovite schist. The graphitic schist predominates in the eastern part of the outcrop. The brown sandy nongraphitic schist at the top of the pre-Silurian section becomes increasingly abundant southward to the essential exclusion south of Blandford Village of the rusty, gray, graphitic schist (see Hatch and Stanley, 1973, p. 7-8 and figure 1). The brown nongraphitic schist is the lower thin-bedded member of the Cobble Mountain Formation of Hatch and Stanley (1973).

An important purpose of this stop is to note the characteristics of the gray graphitic schist of the Middle Ordovician Hawley Formation in order to compare it to the gray graphitic schist of the overlying Silurian and Devonian Goshen Formation to be seen at the next stops.

The open left-handed folds in schistosity have been assigned to Stage III by Hatch (in press). The dominant schistosity is axial surface to the major mappable Stage II isoclinal folds in the Silurian and Devonian Goshen Formation to the east (Hatch, 1968, in press).

Although no indicator minerals higher than garnet are present in this or neighboring outcrops, staurolite and kyanite are both abundant in the Goshen Formation 800 m to the east.

Continue east on Route 20.

0.8 Sanderson Brook Road to right. 150 m to east cross Taconic unconformity and base of Goshen Formation, based on exposures in woods to north and south.

1.4 Turn left onto Old State Road which immediately crosses railroad.

1.6 Cross West Branch Westfield River.

2.1 Enter Blandford quadrangle.

3.9 Outcrop on left is upper sandy thickbedded member of Goshen Formation.

4.4 Footbridge to right over Westfield River.

4.5 STOP 2. West Branch Westfield River, 450 m west of intersection of Fiske Avenue and Basket Street. Park as close as possible along right side of road. Blandford quadrangle (Hatch and Stanley, unpub. data). Outcrop is in river at site of old dam. See figure 2.

Excellent and typical exposure of thinly bedded gray graphitic quartz-muscovite-biotite-garnet-plagioclase-staurolite-kyanite schist and granular schist of the lower thin-bedded member of the Goshen Formation. Rocks here are similar to those in the type area to the north in the Worthington quadrangle (Hatch, 1969) immediately north of the Chester quadrangle. Eastfacing graded beds are well preserved here despite the fact that the

rocks have been metamorphosed to kyanite grade, isoclinally folded (Stage II of Hatch, in press) (see figure 2) and then refolded (Stage III of Hatch, in press) into a major regional fold (figure 2). The northeast-striking northwest-dipping slip cleavage in the more schistose beds is parallel to the axial surface of the later (Stage III) refold.

Note the contrast between this gray graphitic schist and that of the Hawley Formation at Stop 1.

Note the undulating base of many of the graded beds. This structure could be interpreted as primary load casts or as minor folds related to the Stage III slip cleavage.

Note that individual beds here range from a few centimetres to a few tens of centimetres in thickness. Thicker beds are not unusual in this unit to the north, but are rare to the south where individual beds are more typically 1 to 4 cm thick.

Continue southeast along Old State Road toward Huntington Village.

4.7 Bear right at intersection.

4.9 Ridge of thinly bedded Goshen Formation. Graded beds strike N. 80° W., dip 52° N. and top to north.

5.0 Turn hard right by Saint Thomas Church (Huntington Village) onto bridge (Route 112) crossing Westfield River.

5.1 Turn left (east) onto Route 20.

5.5 Enter Woronoco quadrangle.

6.6 Pass through Crescent Mills.

8.7 Village of Russell.

8.8 State Police Parracks on right.

9.0 Outcrop on left opposite Craighurst Gardens is pre-Silurian rocks of the upper member of the Cobble Mountain Formation.

9.2 More upper member of Cobble Mountain on left.

9.5 Cliffs across valley to left are more of same.

9.8 Outcrop on right is more of same.

10.3 Outcrop on right is thinly bedded lower member of the Goshen Formation with boudinaged pods of calc-silicate rock.

10.4 Entrance to Strathmore Park.

10.6 to 11.0 Calc-silicate and sandy quartzose rocks of the upper member of the Goshen Formation.

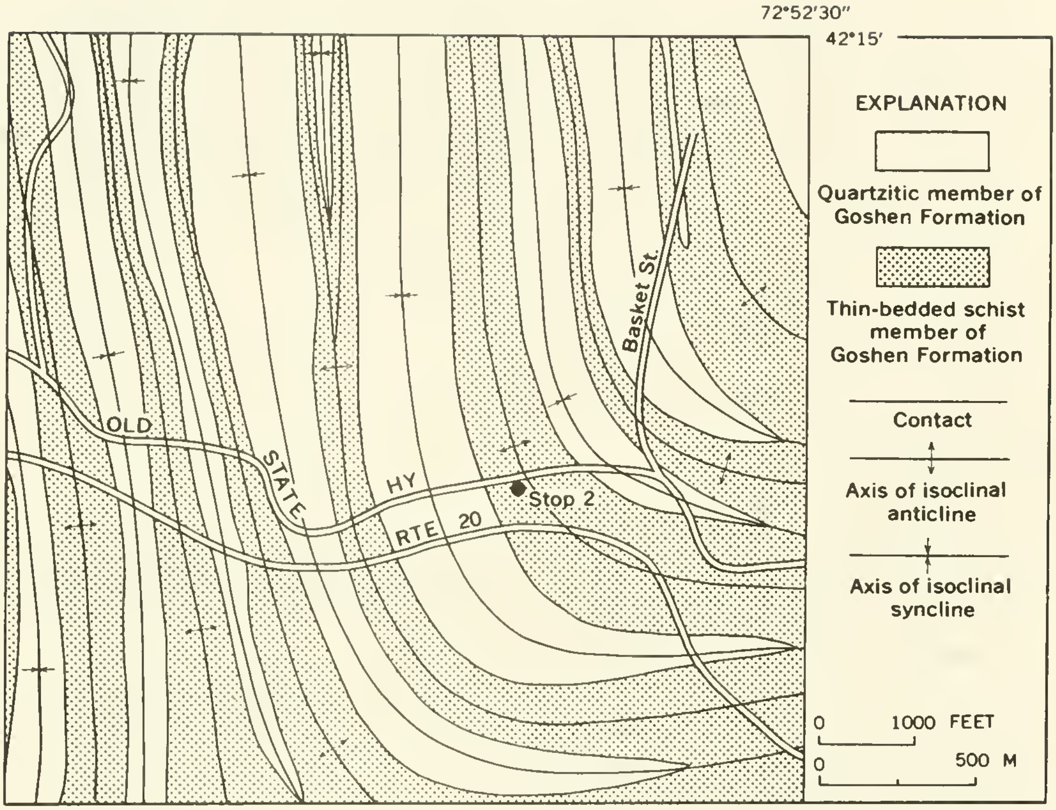

Figure 2. Sketch map of northeast corner of the Blandford, Massachusetts quadrangle showing location of geologic setting of Stop 2.

11.0 Junction of Route 20 and 23. Continue southeast on Route 20.

11.0 to 11.8 Scattered outcrops of Goshen Formation.

11.8 STOP 3. Turn left into Rest Area on north side of Route 20, Woronoco quadrangle (Stanley, S. F. Clark, Jr., and Hatch, unpub. data). Outcrop is on south side of Route 20.

Rock here is thinly bedded quartz-muscovite-biotite-plagioclase-garnet-staurolite graphitic schist typical of the lower member of the Goshen in the southern part of its outcrop area in Massachusetts. Note the pods of calc-silicate rock that we will see more of in later stops. Pegmatites and calc-silicate rocks are boudinaged. The crenulate folds are Stage III of Hatch (in press) and F3 of Stanley (in press). Metamorphic grade has been increasing southward, and fibrous sillimanite has been first recognized about 1.5 km north of here.

Continue southeast on Route 20.

12.1 Pass under Massachusetts Turnpike.

12.2 Enter Triassic.

12.3 Town line, enter Westfield.

12.7 Turn right onto Northwest Road immediately northwest of Four Mile Country Store, and opposite white house with sign "Londys." Immediately bear right at "Y" junction.

12.9 Pass sign for Ralph Lafogg's "Flatstone for sale" on right.

13.6 Turn right on Western Avenue.

A. Bicentennial Note

Two hundred years ago in December 1775, General Henry Knox, in answer to a call from General George Washington who needed artillery to drive the British from Boston, moved 43 cannons and 16 mortars in 40 days from Fort Ticonderoga, N.Y., through Great Barrington and across Massachusetts to Boston over the road now called General Knox Trail on which you are now driving.

13.9 Leave Triassic and enter crystalline rocks.

14.0 STOP 4. Park as directed. Later users of this log can park one or two cars on left opposite outcrop on right (figure 3). Woronoco quadrangle (Stanley, S. F. Clarke, Jr., and Hatch, unpub. data).

Outcrop on north side of Western Avenue (General Knox Road) is pre-Silurian rocks of the upper member of the Cobble Mountain Formation. Although some of these rocks are grossly similar to the rocks of the lower member of the Goshen Formation at Stop 3, they are distinguished from them by the criteria listed in table 1 of Hatch and Stanley (1973) and repeated here as table 1.

The east end of the outcrop consists of graphitic, splintery- and very rusty weathering schist somewhat similar to the graphitic schist of the Hawley Formation at Stop 1. Note that these rocks, although graphitic, weather out into long thin bladelike slivers in a manner characteristic of the Hawley, but not of the Goshen to the north. They are also more fissile and sulfidic than Goshen rocks. Note the minor F3 (Stanley, in press) folds and associated cleavage deforming F2 schistosity at the west end of the outcrop.

From the east end of the outcrop on Western Avenue (General Knox Road) walk due north into the woods for 85 m to outcrop of hard, medium- to fine-grained vitreous to white crystalline quartzite with beds of hard greenish-gray calc-silicate granulite a few metres to the east (figure 3). About 11 m stratigraphically of these rocks are exposed here. We have named this unit the Russell Mountain Formation for nearly continuous exposures from here north to the top of Russell Mountain (about 1.3 km) (Hatch, Stanley, and Clark, 1970). We correlate it with the lithologically and stratigraphically similar Shaw Mountain Formation of Vermont, the nearest exposures of which are about 104 km to the north in Vermont (Doll and others, 1961). Six m to the west are abundant outcrops of the feldspathic rusty-weathering pre-Silurian schist that we just saw on Western Avenue. Fifteen m to the northeast is a small outcrop of gray, graphitic, only slightly rusty, nonsplintery, crinkly quartz-muscovite-biotite-garnet-staurolite-plagioclase schist with 2 mm garnets and the satiny graphitic sheen on fine-grained muscovite surfaces that characterize the schist of the Goshen Formation. The scarcity here of the distinct thin-graded beds that we have seen at Stops 2 and 3 is typical of the basal 30 m of the Goshen and will be noted in subsequent stops. Well-bedded Goshen is abundantly exposed northeast of here.

The open fields to the east are underlain by Triassic sedimentary rocks (fig. 3). Both the Goshen and the Russell Mountain Formations pass under the Triassic rocks immediately south of these exposures and their reappearance or nonreappearance to the south is the controversial subject of the rest of this field trip.

Return to cars and continue northwestward along General Knox Road (continuation of Western Avenue).

Table 1.--Characteristic features of upper Cobble Mountain rocks and basal Goshen rocks useful in distinguishing the two units.

	Upper part of the Cobble Mountain Formation	Basal part of the Goshen Formation
1.	Brown- to orange-brown- to red-rusty-weathering feldspathic and micaceous schist with subordinate beds of nonrusty-weathering feldspathic schist. Commonly nongraphitic except in yellow-rusty-weathering sulfidic zones.	Rusty- to red-rusty-weathering carbonaceous schist generally interlayered with thin (1-10 cm) beds of micaceous quartzite. Where bedded, schist beds are 5-30 cm thick.
2.	Muscovite-biotite ratio commonly 1 in schists; ratio locally greater or less than 1.	Muscovite-biotite ratio commonly greater than 1 in schists.
3.	Muscovite and biotite grains generally same size, producing a coarse "salt and pepper" pattern on some foliation surfaces.	Muscovite grains tend to be larger than biotite.
4.	Garnets in schist 5-15 mm in size. Commonly sparse and cloudy. Where garnets large and numerous, schistosity of rock is bumpy.	Garnets in schist 2-4 mm in size. Commonly more abundant and clearer than in Cobble Mountain rocks.
5.	Graphite, where present, is clotted or forms dark streaks in carbonaceous schists. Graphite commonly absent.	Graphite forms distinctive sheen on schistosity surfaces. Graphite nearly ubiquitous in schists.
6.	Quartz-plagioclase ratio commonly 1 or less.	Quartz-plagioclase ratio greater than 1.
7.	Calc-silicate rocks commonly zoned with quartz, feldspar, garnet, diopside, and epidote abundant in light-colored centers and hornblende concentrated in 2-4 mm zones toward outside of each bed. Calcite absent.	Calc-silicate rocks commonly zoned with thin light-colored centers of quartz, feldspar, garnet, diopside, and epidote and thick rims of hornblende. Calcite commonly present. Calc-silicate rocks generally more hornblende-rich than those in Cobble Mountain Formation.
8.	Plagioclase-quartz-mica gneiss common.	Plagioclase-quartz-mica gneiss rare.
9.	Graded beds very subtle or absent.	Graded beds common and obvious.
10.	Thin (2-6 mm) vitreous quartzites that may be dark colored on fresh surfaces.	Thin, dark-colored, vitreous quartzites absent.

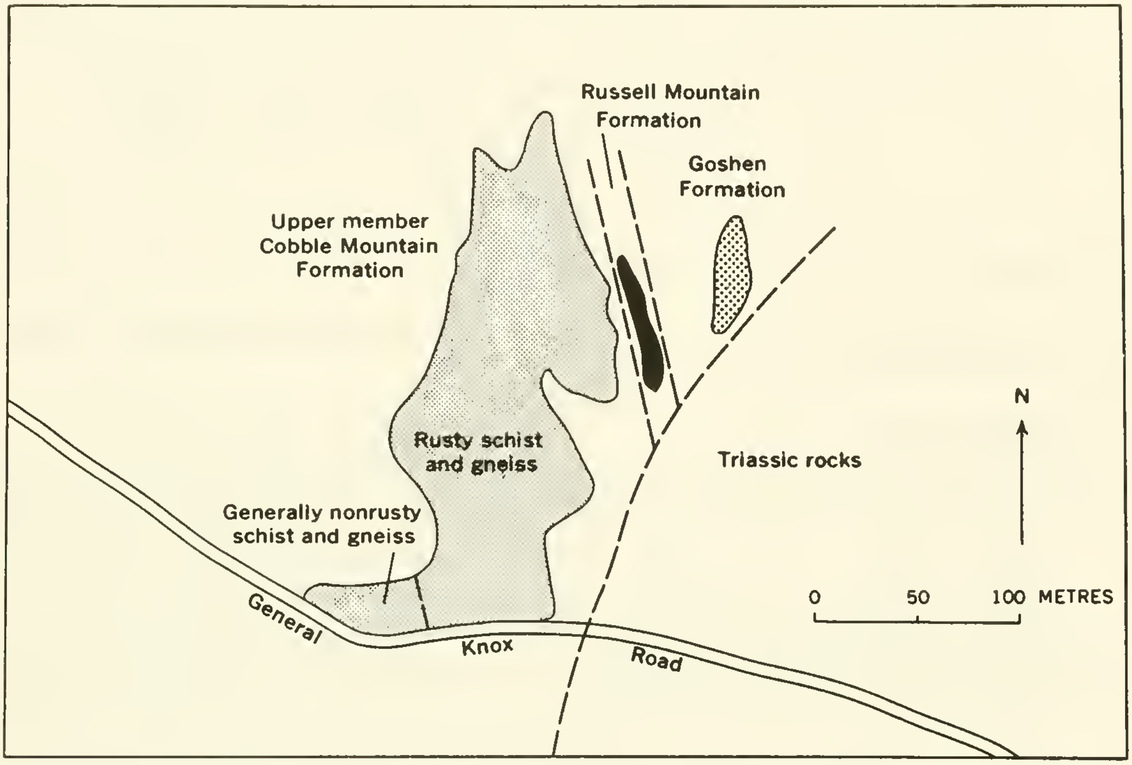

Figure 3. Sketch map of area of Stop 4, southern part of the Woronoco, Massachusetts quadrangle, showing outcrops to be examined at stop.

14.4 Scattered outcrops along road are of upper member of Middle Ordovician Cobble Mountain Formation. Continue northwest on General Knox Road to:

17.5 Turn left (west) onto Route 23.

18.1 Enter Blandford quadrangle.

18.6 Town line, enter Blandford.

19.5 Turn left (south) onto Cobble Mountain Road at red sign for Windy Mountain Farm Sugar House.

19.8 Cross Birch Hill Road, continue south on Cobble Mountain Road.

20.8 Cross Crooks Road, continue south on Cobble Mountain Road.

22.6 Cross Cobble Mountain Reservoir Dam.

22.7 Enter West Granville quadrangle.

23.0 Cross spillway from Cobble Mountain Reservoir and immediately turn into parking area to right.

LUNCH STOP

The rocks along the coast of the reservoir and along the road are the nonrusty and rusty feldspathic schist and gneiss of the thick-bedded upper member of the Cobble Mountain Formation (Hatch and Stanley, 1973, p. 10-12). They are facies equivalents of the black schist and metavolcanic rocks of the Hawley Formation to the north. They are also lithically similar and stratigraphically equivalent to many of the rocks of Stanley's (1964) Collinsville and Rattlesnake Hill Formations to the south in the Collinsville, Conn. quadrangle (Stanley, 1964, 1968; Hatch and Stanley, 1973).

These rocks along the reservoir are deformed by F3 and F4 folds (Stanley, in press).

The mineral assemblage here is quartz-plagioclase-muscovite-biotite-garnet-fibrolite-magnetite. This area and the coast of the reservoir immediately to the west were described in detail by Stanley (1967).

Although the West Granville quadrangle has been mapped by R. W. Schnabel (1973), the interpretation presented for this area is based on independent detailed mapping by Stanley.

Continue south along Cobble Mountain Road.

23.2 Bear right along coast of reservoir.

24.1 Turn left at crest of rise onto Blandford Road (red sign post without sign as of August 1974) and proceed south.

25.0 End of paving; begin dirt road.

25.2 Dirt road enters from right (west); continue south.

27.0 Turn right (west) at "T" onto Route 57 (Main Road).

27.1 Turn left (south) onto unlabelled tar road (Barnard Road on West Granville topo map).

28.3 Unlabelled blacktop road enters from left; continue south.

29.7 Turn right (south) onto East Hartland Road (Route 179).

29.9 State line; welcome to Connecticut.

30.4 Outcrop on both sides of road of thinly bedded graphitic schist of the Straits Schist that we interpret as Goshen Formation re-appeared from beneath the Triassic rocks.

32.4 Enter New Hartford quadrangle.

32.9 Junction with Connecticut Route 20. Immediately beyond turn left (south) onto Route 179 toward Canton, Conn.

36.2 Just before crest of hill turn right into Washington Hill rest area.

STOP 5. Route 179, immediately south of intersection with Hayes Road, Washington Hill. New Hartford quadrangle (R.W. Schnabel, in press). Although Schnabel has mapped this area, the interpretation given here (fig. 4) is ours and does not conform to that of Schnabel.

The north end of the outcrop on the west side of Route 179 is moderately west-dipping thinly laminated light-green calc-silicate granulite. The same rocks crop out on the east side of the road at the crest of the hill. We correlate this rock with the Russell Mountain Formation seen at Stop 4.

South of the calc-silicate rock on the west side of the road is amphibolite and feldspathic schist and gneiss similar to the pre-Silurian rocks at Stop 4 with which we correlate them.

Scattered along both sides of Route 179 for 500 m north of the entrance to the rest area are outcrops of graphitic nonrusty schist of the Straits Schist. Although these rocks here are structurally below the calc-silicate rocks, we believe they are stratigraphically above them and correlative with the Goshen. We attribute the inconspicuous bedding character to the fact that these rocks are at the base of the Straits Schist (Goshen) (see discussion of Stop 4).

We bellieve that the apparent higher feldspar content of the Straits here and to the south relative to the Goshen to the north is due to coarser grain size resulting from slightly higher metamorphic grade and to higher content of intrusive pegmatite and granite.

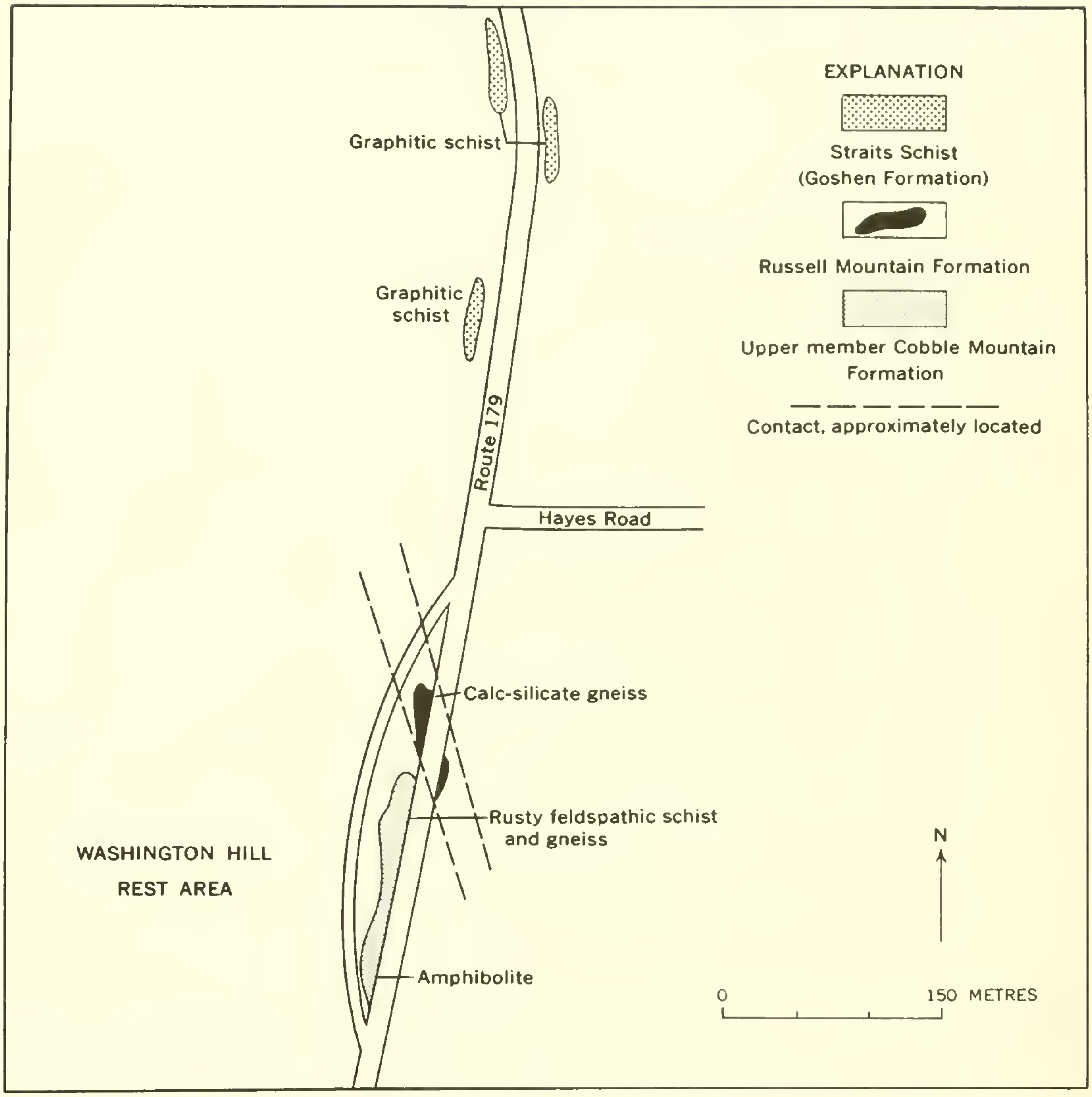

Figure 4. Sketch map of area of Stop 5, New Hartford, Connecticut quadrangle.

We point out that the pre-Silurian rocks differ from the Silurian and Devonian schist here in the following ways:

(1) More biotite and feldspar in pre-Silurian
(2) Less garnet and staurolite in pre-Silurian
(3) No graphitic sheen in pre-Silurian

We suggest comparing hand specimens of the two units.

Return to cars and continue south out south end of rest area and south on Route 179. Outcrops to south along Route 179 to intersection with Route 309 are nonrusty and minor rusty medium-grained feldspathic two-mica schist and gneiss and some volcanic amphibolite that we would correlate with the upper member of the Cobble Mountain Formation.

37.2 Bear left on Route 179 south.

38.7 Outcrop on right is rusty-weathering black sulfide, graphitic schist and black hard vitreous quartzite. Note splintery weathering of schist and fine-grained hard character of quartzite, both reminiscent of the Hawley Formation of Stop 1 and the correlative "upper member" of the Rattlesnake Hill Formation of Stanley (1964) of the Collinsville quadrangle to the south (Hatch and Stanley, 1973).

40.4 Junction with Route 309; continue south on 179.

41.9 Enter Collinsville quadrangle. Outcrops on left are Stanley's (1964) Ratlum Mountain Member of the Satans Kingdom Formation.

42.5 Turn hard left on North Mountain Road.

43.0 Outcrop on left (south) of "lower member" of Rattlesnake Hill Formation of Stanley (1964) (correlative with the upper member of the Cobble Mountain Formation).

43.1 Turn right on East Hill Road.

43.5 Bear right at intersection with Gracey Road.

43.7 Bear right at intersection with Hoffman Road.

44.0 Turn right onto East Mountain Road (Bahre-Johnson Road on Collinsville topo map).

44.7 STOP 6. Between the western part of East Mountain Road (Bahre-Johnson Road on Collinsville quadrangle map) and transmission line 1370 m north-northeast of Rattlesnake Hill. Park 550 m southwest of northern bend in road. Park on the right (west) shoulder of road and pull up as close as possible to sign marked "Slow curve ahead." Traverse is shown in figure 5, Collinsville quadrangle (Stanley, 1964).

Walk 425 m north on East Mountain Road and turn east into pasture through gap in fence. Go north along inside of fence about 15 m to first outcrop of medium- to coarse-grained nonrusty and minor rusty feldspar-biotite-muscovite-quartz-garent-kyanite schist and gneiss with about 1 m of volcanic(?) amphibolite mapped as "lower" member of the Rattlesnake Hill Formation by Stanley (1964). We correlate these rocks with the lithically similar schist and gneiss of the upper member of the Cobble Mountain Formation at the lunch stop. We also correlate these rocks with the somewhat less similar rocks just below (structurally above) the calc-silicate rocks at Stop 5. The folds deforming the schistosity here we call F3 (Stanley, in press).

In contact to the southeast here is 13 m of well-bedded quartzite, calc-silicate quartzite and greenish calc-silicate gneiss indicated by xxx and ccc symbols on Stanley's (1964) Collinsville map at the southeastern contact of his Rattlesnake Hill Formation. We believe these rocks to be the Silurian Russell Mountain Formation seen at Stop 4 (and Stop 5).

Continue walking southeast.

In contact with the calc-silicate-quartzite unit to the southeast is a graphitic rusty-weathering quartz-muscovite-biotite-feldspar-garnet schist having a graphitic sheen on the schistosity. This rock we consider to be the base of the Straits Schist and thus the base of the equivalent Goshen Formation. Once again the base of the Straits Schist is not particularly well bedded. Continue walking southeastward across brook to transmission line to outcrop on transmission line 120 m southwest of East Mountain Road (fig. 5).

Here the rock is well-bedded, thinly-bedded graphitic quartz-muscovite-biotite-garnet-kyanite schist of the Straits Schist that we consider to be Goshen. Note that the rock does not have the splintery-weathered texture, fine-grained clean quartzites, or very rusty weathering that characterize the pre-Silurian graphitic schist.

Graded beds present in outcrop top southeastward and are compatible with, but do not necessarily prove, because of the break in outcrop immediately to the west, our stratigraphic interpretation that Straits is younger than the rocks to the northwest.

Another outcrop of Goshenlike Straits Schist with thin beds similar to that at Stop 3 is 150 m to the southwest along the transmission line. Pods of mineralogically zoned calc-silicate amphibolite are very similar to those in the Goshen Formation to the north.

Note pods of quartz and blue kyanite.

44.7 Continue south then west down East Mountain Road.

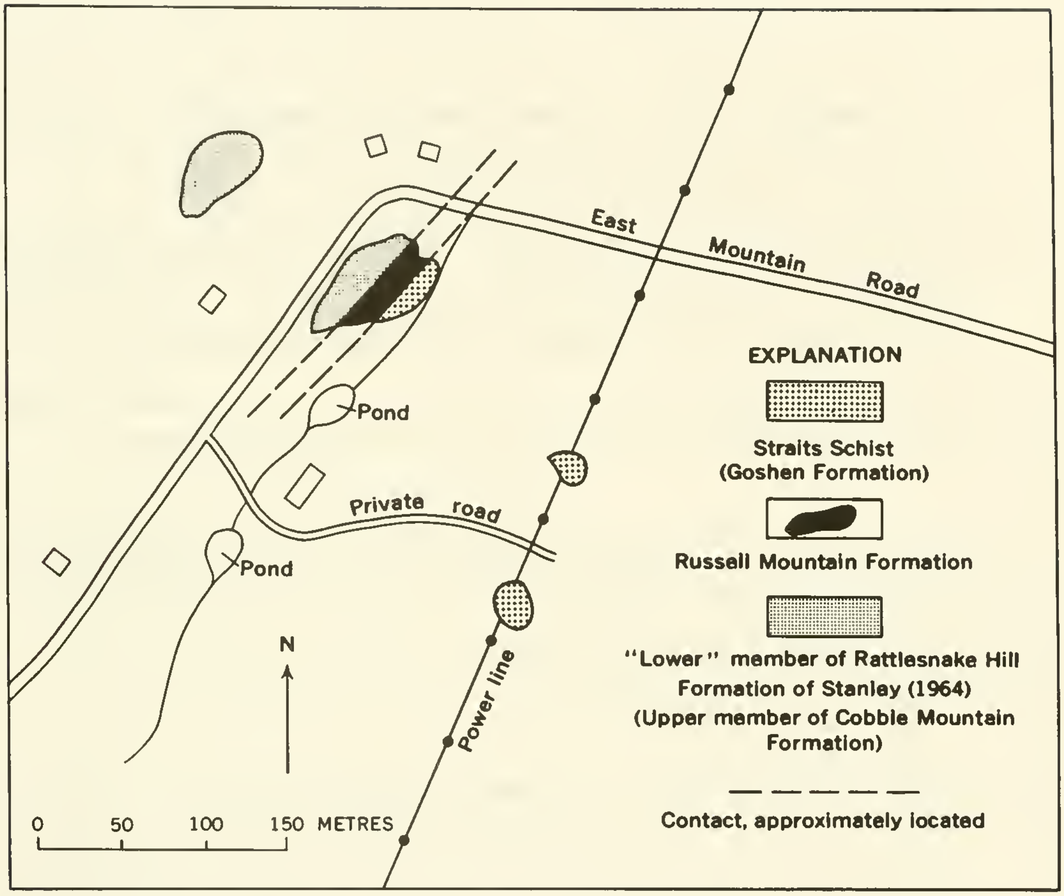

Figure 5. Sketch map of area of Stop 6, Collinsville, Connecticut quadrangle.

45.2 Turn left (south) on Route 179 at "T."

46.5 Junction Route 179 and Route 44. Go south to Collinsville on Route 179.

46.7 Junction Route 202. Continue south on Route 179.

48.0 Junction. Turn right at "Y."

48.1 Bear right along Farmington River. Stay on Route 179.

48.4 Cross Farmington River. Turn left (south) staying on Route 179.

48.6 Begin large roadcut of Middle Ordovician Collinsville Formation (of Stanley, 1964). Northern part consists mainly of amphibolite. To the south plagioclase gneiss and some beds of coticule (fine-grained, pink quartz-garnet granulite) are abundant.

48.7 Feldspathic schist of Stanley's (1964) Sweetheart Mountain Member of the Collinsville Formation. South of the end of the large exposure is a small outcrop of Straits Schist. Equivalents of the Russell Mountain Formation are not present here.

49.1 to 49.3 Large exposure of the Straits Schist along the eastern limb of the syncline between the Collinsville and Bristol domes.

50.4 Turn right (west) onto Route 4 (Route 116 on 1956 edition of the Collinsville quadrangle) passing west through the syncline between the domes.

50.6 Enter the Bristol dome.

52.7 Enter Burlington, Conn.

53.0 Junction Route 69 (not so shown on 1956 edition of Collinsville quadrangle). Continue west on Route 4.

53.1 Turn right into Woodruff Hill Rest Area behind long road cut through the Straits Schist on the western side of the Bristol dome.

<u>STOP 7</u>. Roadcut along Route 4, 150 m west of junction with Route 69. Park in Rest Area to north marked Woodruff Hill (Collinsville quadrangle, Stanley, 1964).

Outcrop consists of medium-grained quartz-muscovite-biotite-feldspar-garnet-kyanite schist with graphitic sheen. Beds of quartz-biotite-garnet granular schist are interlayered with schist and are typical of central part of this belt of Straits Schist. Mineralogy, bedding character, and graphitic sheen are identical to the Goshen Formation to the north. Some granular beds are graded with tops to the southeast. Several beds of calc-silicate amphibolite similar to those at Stops 3 and 6 are boudined. F3 folds deform schistosity which cuts bedding at a slight angle indicating structural tops (synclinal axis) to the southeast. Continue west on Route 4 to Harwington in the Torrington quadrangle.

56.1 Junction Route 72 and Route 4. Continue west on Route 4.

57.1 Harwington-Harmony Road to the right (north).

57.9 Center of Harwington (Harwington Green).

58.1 Junction Route 118 and Route 4. Continue west on Route 118. Outcrop to left (south) is Hartland III of Martin (1970); equivalent to Ratlum Mountain Member of Satans Kingdom Formation (Moretown Formation) and the "upper" member of Rattlesnake Hill Formation (Hawley Formation) of Stanley (1964).

59.1 Junction Route 222 and Route 118. Continue west on Route 118.

60.7 Turn left on Route 8 south toward Reynolds Bridge and Waterbury.

68.6 Leave Route 8 at Exit 38 (Route 6 to Watertown). Outcrop to the south of exit shows beautiful recumbent folds in the Reynolds Bridge Gneiss (Collinsville Formation as used by Stanley, 1964) of Cassie (1965).

Continue southwest on Route 6 toward Black Rock State Park and Watertown.

69.6 Black Rock State Park.

69.9 Bear right onto Bidwell Hill Road.

70.5 Outcrops from here to the southern intersection of Bidwell Hill Road and Route 6 are well-bedded feldspar-biotite gneiss and schist with minor volcanic(?) amphibolite. These rocks are mapped by Robert Cassie (1965) as Reynolds Bridge Gneiss which we would correlate with the Collinsville Formation as used by Stanley (1964) and place at the top of the pre-Silurian section.

70.9 Turn 180° left (north) onto Route 6. Dangerous turn. Drive slowly on right side of road so as to park safely near yellow sign warning of a side road (Park Road) entering on the right.

STOP 8. Route 6 just east of Bidwell Hill Road at the junction of Route 6 and Park Road (fig. 6). Thomaston quadrangle, Cassie (1965).

Walk 110 m south from sign along east side of road to first outcrop. Here the rock is a medium- to coarse-grained feldspar-biotite-quartz gneiss typical of Cassie's (1965) Reynolds Bridge Gneiss and Stanley's (1964) Collinsville Formation. Although amphibolite is not visible in this outcrop, it is present on Bidwell Hill Road to the west as well as in an outcrop just north of the junction of Route 6 and Park Road (fig. 6).

Walk north 130 m from yellow sign to the junction of Park Road and Route 6. On the north corner is a small outcrop of hornblende-plagioclase amphibolite of the Cassie's Reynolds Bridge Gneiss. This can be traced northwesterly across Route 6 to the outcrop just east of the gray house (fig. 6). In contact with the amphib-

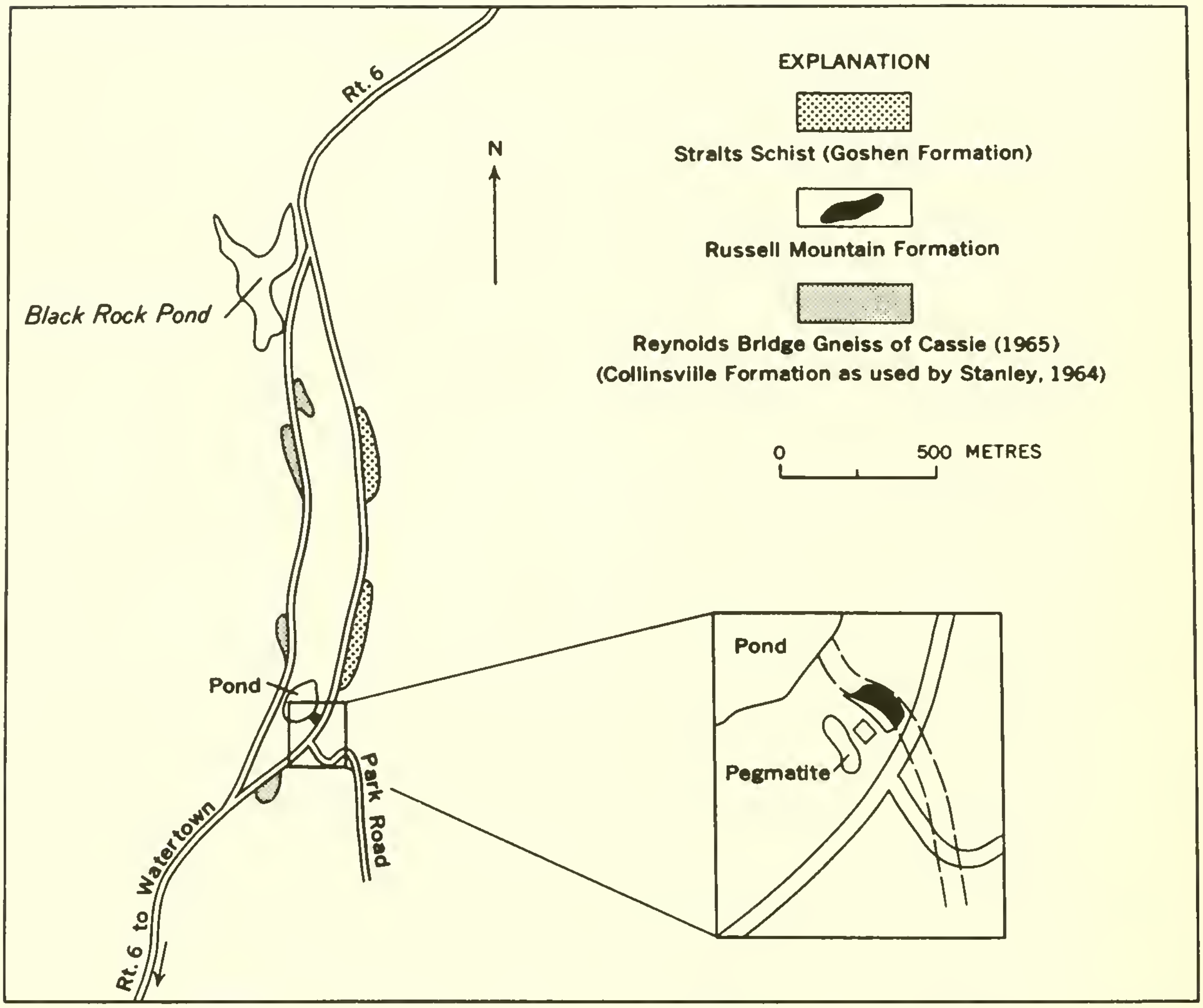

Figure 6. Sketch map of area of Stop 8, Thomaston, Connecticut quadrangle.

olite to the northeast is approximately 3 m of well-bedded marble and calc-silicate quartzite. We believe this calc-silicate unit is the Russell Mountain Formation and, therefore, stratigraphically above, though here structurally below, the amphibolite and feldspar gneiss of Cassie's Reynolds Bridge Gneiss.

Continue walking north along Route 6 about 130 m to a long outcrop of quartz-muscovite-biotite-garnet-feldspar-kyanite schist with satiny sheen of fine-grained muscovite and graphite. This is the Straits Schist of Cassie (1965). We concur and correlate this with the Goshen Formation of western Massachusetts.

Thus we have walked over the same section from pre-Silurian rocks, through the Silurian calc-silicate-quartzite of the Russell Mountain Formation, to the bedded graphitic quartz-muscovite schist and granular schist of the Goshen Formation or correlative Straits Schist that we saw at Stop 4, Stop 5, and Stop 6.

As seen on the earlier stops, the basal part of the Goshen and correlative Straits Schist is only indistinctly bedded, but it grades upward into well-bedded rock like that at Stop 7.

Note that the grain size of the Straits is coarser here than at Stop 7. This southward increase in grain size continues into the Naugatuck quadrangle where the Straits is conspicuously coarser than here.

END OF TRIP. We thank you for bearing with us and hearing our story.

71.2 Continue north on Route 6 past Black Rock State Park.

72.8 Bear right, follow Route 6 east to Route 8.

73.3 Turn onto Route 8 north. Take Route 8 to Winsted, then Route 44 northwest to Route 7 and then Route 7 to Great Barrington.

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TRIP B-4 (Sat.) and C-4 (Sun.)

Stratigraphic and structural relationships along the east side of the Berkshire massif, Massachusetts

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Introduction

The purpose of this trip is to examine the evidence for: (1) polyphase deformation and metamorphism of Precambrian and Lower Cambrian rocks at the north end of the Berkshire massif and southward for 30 miles along the east side; and (2) premetamorphic thrust faults which have produced an interlayering of Lower Cambrian schists and Precambrian gneisses and plutons.

Stratigraphy

Gneisses of the Berkshire massif are composed dominantly of quartz, plagioclase, and microcline[1] and accessory amounts of garnet, various amphiboles, biotite, or rare muscovite. Metapelites, quartzite, calc-silicate gneiss, and marble are subordinate regionally but locally are common. Openly to tightly folded thrust slices make regional correlation of less distinctive lithologies difficult. All these rocks are believed to be Precambrian in age. Whole-rock Rb/Sr ages on the gneisses (Brookins and Norton, 1975) suggest an age of about 1.1 b.y. These lithologies are intruded by igneous rocks of two separate ages, 0.9 and 0.6 b.y. (Brookins and Norton, 1975).

At the north end of and along the east side of the massif (Fig. 1), the Precambrian rocks are in contact with aluminous rocks of the Hoosac Formation of Early Cambrian or older age. The detailed stratigraphy has been discussed by Norton (in press). The rocks range from quartz-albite-muscovite-chlorite (or biotite at higher grade) to quartz-muscovite-garnet ($\pm$ staurolite, kyanite, and sillimanite) schist.

The foliation of the rocks of the massif and the Hoosac schists is nearly everywhere parallel at their mutual contact. This has led some workers to interpret the contact as depositional (Gates et al., 1973). Emerson (1898) interpreted part of the "Becket Granite Gneiss" at the base of the Hoosac as a conglomerate, the contact between the two formations being a profound unconformity. These rocks have been reinterpreted as pre- and synmetamorphic tectonic breccias and mylonites by Norton (1975). The contact between the Precambrian rocks (Brookins and Norton, 1975) and the Lower Cambrian or older Hoosac rocks along the entire eastern margin of the

[1]Minerals are listed in approximate order of decreasing abundance

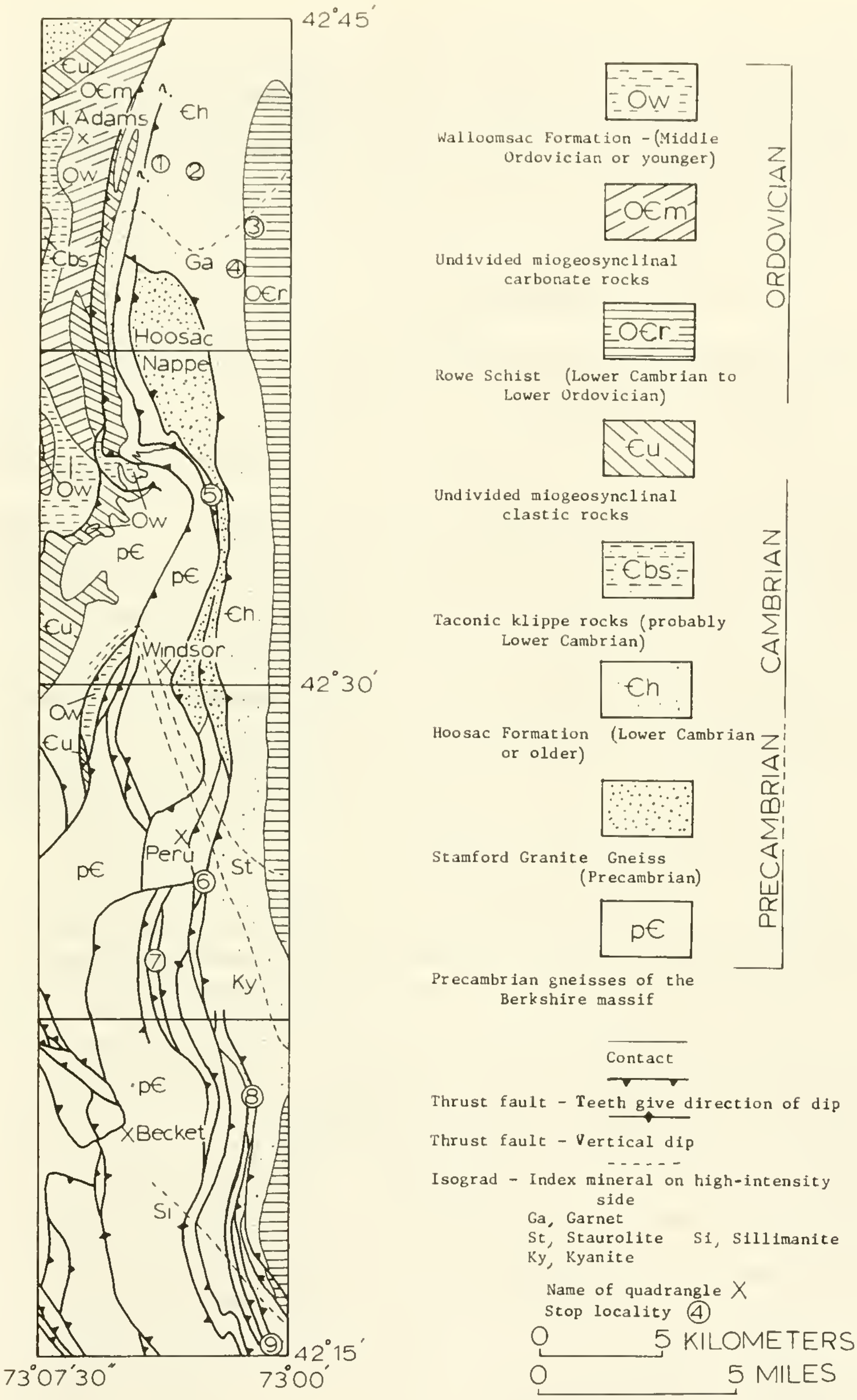

Figure 1: Geologic map of the east margin of the Berkshire massif, Massachusetts. (Geology of the North Adams quadrangle generalized from Herz, 1961.)

massif is interpreted to be a premetamorphic fault.

Directly overlying the Hoosac to the east of the massif is the Rowe Schist (Hatch et al., 1966) of Early Cambrian to Early Ordovician age. The Rowe consists primarily of unbedded quartz-muscovite-chlorite schist and accessory garnet, magnetite, and plagioclase. Quartz lenses and ribbons are common. The contact with the Hoosac is gradational. The Rowe will be seen only at STOP 3.

Structure

At least six deformations have affected the rocks at the north end of the Berkshire massif. Details of these events have been given by Norton (1975). Briefly, the different episodes are as follows:

F0: This deformation is evidenced by isoclinal folds and related penetrative foliation, found only in rocks of Precambrian age.

F1: The dominant structures west and north of the massif consist of isoclinal folds and axial-plane schistosity. Attitudes of these folds vary systematically from recumbent, gently north-plunging in the west (STOPS 1 and 2) to steeply east-dipping with variable plunges in the east (STOPS 3, 4, 6, 8 and 9). The age of this deformation is believed to be Ordovician. Concurrent with formation of small-scale folds are large-scale recumbent folds and soft- and hard-rock thrusts involving Paleozoic and Precambrian rocks, respectively. Evidence viewed at STOPS 1 and 2 suggests that F1 may consist of two separate events.

F2: The intensity and orientation of F2 structures vary systematically from west to east. On the west (STOPS 1 and 2), F2 folds are open, north-striking, 45° east-dipping and shallowly plunging. No axial-plane foliation has formed. Eastward, F2 becomes more intense, and the dip of foliation steepens to nearly vertical, with plunges nearly down the dip. At STOPS 3 and 4, F1 and F2 are of nearly equal intensity. Southward, F2 is more intense than and commonly obliterates F1 structures. However, all F2 structures appear to be intraformational. F2 deformation is believed to be related to the Acadian orogeny. An episode of recumbent folding in rocks of Silurian and Devonian age (Osberg, 1975) which is older than F2 is not present in rocks of Precambrian, Cambrian, and Ordovician age. This suggests the existence of a décollement at the Ordovician-Silurian boundary.

F3: F3 folds are open to closed and strike northeast. Axial planes dip steeply northwest, are vertical, or dip steeply southeast. Axes plunge generally northeast. These folds are not observed west of the axis of the massif and become more intense to the southeast along the east side of the massif. F3 is believed to be an Acadian deformation.

F4: F4 is characterized by the formation of north-striking warps and crinkles on F1 and F2 foliation surfaces. No foliation or metamorphism is associated with this deformation. The age of F4 is unknown, but it probably is Devonian to Triassic in age.

F5: Rarely one observes small postmetamorphic faults that cut F1, F2, and F3 structures. F5 probably postdates F4 and may be Triassic in age.

Metamorphism

The regional isograds are shown on Fig. 1. These isograds integrate the effects of at least three Paleozoic events (M1, M2, M3). Details of the metamorphic history of the area are given by Norton (1975).

The earliest recorded metamorphic event (M0) at the north end of the massif is recorded in the Precambrian rocks, which have not been severely retrograded, or overprinted with a high-grade Paleozoic event, or reconstituted by Paleozoic folding and premetamorphic faulting. Textures and mineral assemblages in calcareous gneisses and metapelites suggest that M0 was at least of sillimanite grade.

M1 spanned F1 in time. The thermal peak (garnet grade) occurred slightly after F1 folding and faulting. The locus of the most intense M1 metamorphism appears to be west of the axis of the massif.

M2 spanned F2 in time, with a thermal peak at garnet grade east of the axis of the massif.

M3 preceded F3 in the area nearest to the Precambrian-Cambrian contact at the east side of the massif. Further east, in rocks of Silurian and Devonian age, porphyroblasts of kyanite and staurolite crosscut F3 foliation suggesting that M3 was time transgressive to the east. Within the Hoosac, the production of staurolite and kyanite appears to be an M3 event.

The retrograde metamorphism of the Precambrian gneisses most probably was accomplished during M1. No Silurian and Devonian rocks are in contact with the Precambrian gneisses. Consequently, the Cambrian and Ordovician rocks that provided water for the retrogression of the gneisses during M1 would in turn act as a sink for infiltration of water from younger rocks during subsequent metamorphisms.

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Road Log

Outcrops will be visited in the North Adams (Herz, 1961), Windsor, Peru, and Becket quadrangles, Massachusetts (1:24,000). The road log begins at the entrance of the Monument Mountain High School, Great Barrington, Massachusetts. The total mileage is 111 miles. Gas up before the trip.

00.0	Start of trip. Allow 1 hour for travel to assembly point in North Adams. Proceed north on Route 7. Departure from North Adams will be at 8:30 a.m. on Saturday. Sunday trip departs from the Monument Mountain High School parking lot at 7:30 a.m.
02.4	Stockbridge. Turn right on Route 7.
04.1	Pass under Massachusetts Turnpike.
15.5	Pittsfield center. Turn right. Follow signs for Route 7 and 9.
15.7	Left on Route 7 and 9.
16.4	Right on Route 9.
17.1	Bear left on Route 9.
18.9	Left on Route 8 by "MacDonalds".
22.3	Cheshire town line (type locality for the Cheshire Quartzite). View of Taconic Klippe on Cambrian and Ordovician carbonate rocks.
26.2	Cheshire center.
31.2	Intersection of Route 8 and 116. Continue on Route 8 (many gravel pits from Cheshire to North Adams are in kames).
34.5	Turn right on Route 8A (poorly marked). Hill directly ahead has, in inverted stratigraphic order, Hoosac Formation on Dalton Formation on Cheshire Quartzite at the base of the hill. The first bench is a kame terrace, which can be traced for 6 miles.
37.6	Intersection in North Adams. Turn right on Main Street (Route 8A signs have disappeared).
37.7	Turn left on Church Street by monument.
37.8	Turn right on Routes 2 and 8.
38.5	Fork right on Route 2.
39.6	View of west side of the Hoosac Range. "Flat-lying" but overturned Dalton Formation at base, Hoosac Formation above. The

regional strike is parallel to the ridge.

39.8 Power line - reassemble. Reset mileage to 00.0.

00.0 Start

00.4 Outcrop of gently west-dipping Dalton Formation overlain (not visible) by Hoosac Formation.

01.6 Hairpin turn. Continuous outcrop of Hoosac Formation for next mile.

02.3 STOP 1: Pull off to right in parking area. Be careful of traffic!

Rocks typical of the Hoosac Formation are exposed continuously for about a mile on the east side of the road. The dominant lithology is a quartz-albite-muscovite-biotite $\pm$ chlorite schist. Albite is typically porphyroblastic and black, containing appreciable amounts of graphite inclusions. The garnet isograd is an unknown distance to the southeast (three miles at the most). A few tan to gray fine-grained quartzite beds are very discontinuous. Many ribbons and lenses of white vein (?) quartz outline what appear to be recumbent isoclinal folds (F1) which plunge gently to the north. Small-scale kink folds (F2) without associated foliation are rarely present. More commonly, small faults have formed parallel to the north-striking axial planes of F2 folds. Axial surfaces dip about 45°E. Axes are subhorizontal. These faults commonly are marked by vein quartz. The quartz ribbons outlining F1 folds are interpreted to have formed early in F1 and were followed by the formation of a penetrative axial surface schistosity and refolding of the ribbons (see Raybould, 1975, for a supporting argument).

Continue east on Route 2.

02.5 Lookout. Taconic klippe to the west underlain in the valley by Cambrian and Ordovician carbonate rocks.

03.1 STOP 2: Park on right shoulder of Route 2. Be careful of traffic! Outcrop is on north side of road opposite "Bob's Variety Store". NO HAMMERS PLEASE!

The lithology exposed here is atypical of the Hoosac Formation. It has considerably less muscovite and ferromagnesian minerals than normal. Typical assemblages are quartz-albite-muscovite-biotite $\pm$ chlorite $\pm$ magnetite. The typical porphyroblastic albite is present. The well-displayed compositional banding in the sandier lithologies suggests bedding, and a graded bed or two are present. F1 isoclinal recumbent folds are outlined by quartz lenses and ribbons. F2 folds are conspicuous, commonly having an axial surface cleavage and axial surface quartz veins. The orientation of F2 structures is similar to that at STOP 1.

Continue east on Route 2.

04.8 Outcrop on left of Hoosac Formation, continuous outcrop for 0.2 miles to

05.0 Rowe Schist. Whitcomb Summit. Within this outcrop, F1 and F2 are nearly equal in intensity. F3 and F4 folds are unsystematic and cannot be differentiated with certainty. F1 foliation gets progressively steeper to the east. Here F2 axial surfaces dip about 60°E. To the east, across the Deerfield Valley one can see the Bear Swamp pumped storage facility which is entirely within the Hoosac Formation (Chidester et al., 1967).

05.9 Pass over the Hoosac Tunnel (1,000 feet below) (see Pumpelly et al., 1894, for the geology of the tunnel). All outcrops from Whitcomb Summit to STOP 3 are in the Rowe Schist.

06.3 STOP 3: "Eastern Summit." Park on right shoulder of the road in front of cabins. Pavement outcrop between road and cabins.

The typical Rowe Schist exposed here is a quartz-muscovite-chlorite schist and accessory magnetite and rare garnet. Here, F1 and F2 foliations (the latter is most commonly a shear foliation in this area) are about equally developed. The intersection of the nearly parallel foliations produces an anastomosing fabric that gives the rock a rhombic patterned surface. F3 kink folds are present locally. Pleistocene features include well-developed glacial grooves and striations which are parallel to the topographic contours and the regional ice-movement direction. Additionally there are striations more or less parallel to the slope. Five suggestions are offered for their origin (take your choice):

(1) Downhill creep of till.
(2) Multiple glaciation.
(3) Ice disintegration flowlines perpendicular to the slope (T. Hughes, personal communication, 1975).
(4) Man's activities.
(5) All of the above (D.S. Harwood, personal communication, 1975).

Continue east on Route 2.

06.6 Turn right on Church Road (you are now in Florida!).

07.4 Turn right on South County Road.

08.0 STOP 4: Pavement outcrop on the north side of the road and outcrops to the northeast. Sign points to foot trail to Whitcomb Summit. NO HAMMERS PLEASE! (This outcrop may be omitted if the group is large or if we are running behind schedule).

The outcrop is near the top of the Hoosac Formation. The rock is a quartz-albite-muscovite-chlorite ± magnetite schist. Near the

road, the intersection of F1 and F2 foliations give the same anastomosing texture seen at STOP 3. In the woods at the north end of the outcrop is a small, postmetamorphic open upright fold interpreted to be F4. F3 folds are rare in the pavement outcrop but common in the outcrop about 150 feet to the northeast (also in the Hoosac).

Continue west on South County Road.

09.4 Turn left on Savoy Road (becomes New State Road). You are now within the Stamford Granite Gneiss of Herz (1961). All exposures between this point and STOP 5 are in this unit.

12.3 Keep left; Burnett Road forks right.

13.8 Turn left on Adams Road.

14.0 Turn right (Savoy Center) on Center Road.

16.9 Intersection with Route 116. Turn right.

17.1 Cross Savoy Hollow Brook, start of long road cut.

STOP 5: Roadcut on both sides of Route 116. Pull well to right.

(This outcrop is known privately as the type locality of the "Cuddly Bunny Formation").

The easternmost part of the outcrop consists of about 50 feet of cataclastic microcline augen gneiss. This rock is equivalent to the Stamford Granite Gneiss mapped by Herz (1961) in the southern part of the North Adams quadrangle and may be equivalent to the type Stamford. The contact with the Hoosac to the east is interpreted to be a premetamorphic fault zone (F1). West of the augen gneiss is a section of plagioclase-quartz-biotite gneisses, locally folded and faulted by postmetamorphic movement (F3?) (F5?). Garnetiferous schist of the Hoosac Formation is present in the middle of the roadcut. The septum, bounded on both sides by mylonite, may be traced northward and westward along the margin of one of the major thrust slices. Further west in the roadcut are well developed breccias and mylonites in Precambrian gneiss (Figs. 2 and 3).

Continue west on Route 116.

17.3 Left on Route 8A.

21.8 Intersection with Route 9. Turn left.

21.9 Turn right on Peru Road, just before school.

22.4 Bear left.

Figure 2: Pseudoconglomerate formed by the partial transposition of compositional banding (FO) in Precambrian gneisses by F1 faulting. STOP 5: Scale is 7 inches long.

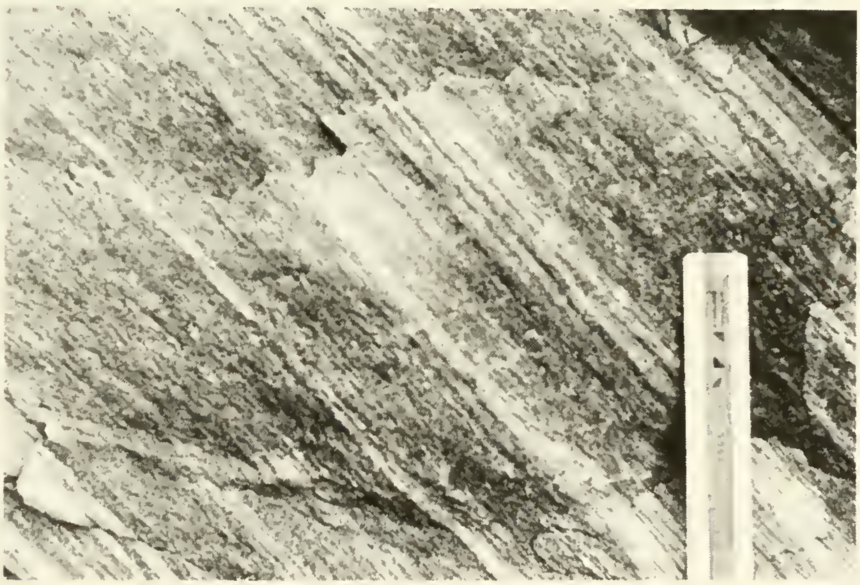

Figure 3: Mylonite formed within the gneiss near the Precambrian gneiss-Hoosac contact. STOP 5: Scale is 7 inches long.

24.5 Pass under powerline. Crop to east (left) of road is Hoosac schist within the Precambrian terrane.

25.0 Town line. Entering Peru. Road becomes Beauman Road.

26.1 Turn right on West Windsor Road.

26.5 Bear right.

28.0 Bear left on dirt road (North Road).

29.2 Route 143. Peru Center. Continue south across Route 143 on South Road.

30.0 Left at "T" junction.

30.2 Entrance to Dorothy Francis Rice Wildlife Sanctuary. Continue east on dirt road.

30.6 Parking area for visitors.

STOP 6: Park in field just before visitors' information center. Walk east on footpath, through swale, cross the wall, and stop at the cliffs on the north side of the trail -- about 1/4 mile from the information center.

Note the NE. strike (and NW. dip) of the Precambrian gneiss as you walk in. At the contact of the Precambrian gneiss with the Hoosac Formation, the gneiss is reoriented by folding and cataclasis so as to parallel the foliation within the Hoosac. Near the south end of the outcrop is slight angularity between the schist and gneiss. The gneiss is composed dominantly of quartz + plagioclase + microcline + biotite. The Hoosac is a quartz-albite-muscovite-biotite schist. F1 and F2 are indistinguishable here. One thin tectonic sliver of gneiss is found within the schist. F3 structures are well developed. They occur as open folds in the schist and pass into quartz-filled fractures in the gneiss.

31.0 Return to paved road and turn left (south) on South Road.

31.6 Phone line crossing.

32.0 Turn right on South Road.

32.5 Intersection with Middlefield Road. Turn left.

33.1 STOP 7: Pull off to the right shoulder opposite the green house on the east side of the road (The SKY LINE CLUB). Traverse through the woods to the west for 1,500 feet.

Several lithologic variations of Precambrian gneiss are present. From 1,000 feet to 1,500 feet, we will pass through a zone of

extensive brecciation and mylonitization with a few septa of schist, interpreted to be Hoosac schist. This particular zone has been followed for about 10 miles to the south. The zone generally parallels the Precambrian-Hoosac boundary at the east margin of the massif but locally truncates Precambrian stratigraphic units.

Continue south on Middlefield Road.

34.0 Town line. You are now in Middlefield on Main Road.

34.1 Prominent ridge on left is capped by the lower units of the Hoosac. The Precambrian-Cambrian contact is at the break in slope at the base of the ridge.

37.9 Town Hill Road to right, Middlefield Center.

STOP 8: Pull off to right opposite Mobil Station. Traverse will start just south of the town hall where low outcrops of well-banded contorted gneiss (interpreted to be Precambrian) are exposed. This band of gneiss noses out within 500 feet to the north but widens to more than 200 feet within 1,000 feet to the south. The contacts between the gneiss and bounding schist are razor sharp. Walking westward, one goes through a highly aluminous quartz-muscovite-garnet-biotite-staurolite $\pm$ kyanite schist, typical schist of the bulk of the Hoosac (quartz-muscovite-garnet-biotite), more aluminous schist (quartz-muscovite-garnet-biotite), typical albite schist, and finally into cataclased quartz-plagioclase-microcline gneiss with conspicuous magnetite octahedra. The magnetite postdates the cataclastic foliation and is responsible for prominent magnetic anomalies of the area. This contact is interpreted to be a major premetamorphic thrust fault (also seen at STOP 6), which extends from the north to the south end of the Berkshire massif, along the east margin. F1 is not distinguishable here. The isoclinal folds outlined by quartz ribbons are interpreted to be F2 in age. F3 folds contort the schistosity and have cracked the garnet porphyroblasts. Meta-igneous rocks from a drill hole 4,000 feet west of this locality yielded a Rb/Sr age of 895 m.y. (Brookins and Norton, 1975).

38.0 Turn around and head west on Town Hill Road.

39.2 Cross Factory Brook. Colesbrook Limestone (Precambrian) is exposed here and has been traced for approximately 7 miles parallel to the Paleozoic regional strike and parallel to the faults in the area.

41.2 Cross Factory Brook.

41.4 Crop on left. Hoosac structurally <u>overlain</u> to the east by Precambrian gneiss.

41.7 Bridge arch constructed from "Becket Granite Gneiss" (Emerson, 1899).

41.9 Crop on right. Hoosac in contact to east with Colesbrook Limestone. Compare with mile 41.4, 1,200 feet northeast along strike.

42.1 Cross the West Branch of the Westfield River. Now on Bancroft Road.

47.8 Crop on left, starts in Precambrian rocks, ends in Hoosac. Rocks on both sides of the contact are structurally conformable.

47.9 Route 20. Turn right.

48.0 Cross Walker Brook. The Precambrian-Cambrian contact is exposed in the brook at several localities.

48.4 Hard left on Quarry Road.

49.3 Gates across dirt road to left (on North side of small hill).

STOP 9: Park well on right off road. Walk southeast 2,200 feet on dirt road on quarry in quartz monzonite.

Flaggy well-bedded quartz-plagioclase-biotite-muscovite gneiss of the Hoosac Formation is in sharp contact with weakly foliated quartz monzonite. The quartz monzonite contains no inclusions of Hoosac nor does it crosscut it. The development of foliation in the quartz monzonite is greatest at the contact. These relationships suggest that the Hoosac and quartz monzonite are in fault contact. A Rb/Sr age of 605 m.y. (Brookins and Norton, 1975) and a zircon date of 760 m.y. (Zartman, oral communication, 1973) suggest that the upper Precambrian plutonic rock may be nearly concurrent with the commencement of the deposition of the Hoosac. Later faulting (Taconic) placed them in contact.

Turn around and return to Route 20.

50.2 Left on Route 20.

52.0 Route 8 goes right. Continue west on Route 20 and 8.

57.2 Route 8 goes left. Continue west on Route 20.

64.1 Left on Route 102, just before toll booth for Massachusetts Turnpike.

66.8 East Lee.

68.5 Route 102 joins Route 7. Continue on Route 7.

68.9 Center of Stockbridge. Red Lion Inn. Follow Route 7 to the left.

71.3 Monument Mountain High School.

THE CAMBRIAN-PRECAMBRIAN CONTACT IN NORTHWESTERN CONNECTICUT AND WEST-CENTRAL MASSACHUSETTS

Robert W. Schnabel[1], Charles W. Martin[2], and Robert M. Gates[3]

Introduction

Mapping in the West Granville, Tolland Center, Winsted, Torrington, West Torrington, Cornwall and South Canaan quadrangles shows that the contact between so-called Precambrian rocks to the west, and Cambrian rocks to the east is not sharp, clear cut and marked by a major fault or an unconformity. Rather, rocks with lithologies typical of Precambrian rocks exposed in the Berkshire and Housatonic massifs to the west, and rocks with lithologies typical of Cambrian lithologies to the east, are interlayered at many different scales throughout the area. This area is along the eastern flank of the Berkshire and Housatonic Highlands Massifs, along the border between Connecticut and Massachusetts (Figure 1).

Distribution of the geologic units as they have been mapped within the area are shown on figure 1 which is a much simplified version of our geologic maps.

Terminology of the formations within this area has undergone an extensive and anguished evolution. The gneisses have been called "Becket Gneiss", "Stamford Granite Gneiss", "Hinsdale Gneiss", "Washington Blue Quartz Gneiss", "Gneiss complex of the Berkshire Highlands","Gneiss complex of the Housatonic Highlands", as well as other less formal names. Most often the schist units have been assigned to the Hoosac Schist in Massachusetts, and to the Waramaug Formation in Connecticut. For the purposes of the present report, we shall use the terms "schist" and "gneiss" and beg the questions of precise stratigraphy.

Characteristically the gneisses are hard, dense rocks that tend to break rather erratically, about as easily across the foliation planes as parallel to them. They are mostly quartz, plagioclase, microcline, biotite rocks with relatively little muscovite, in fact many of the gneisses contain no muscovite at all. Accessory minerals are not abundant in most of the gneisses, they tend to be small and widely dispersed through the rock where they do occur.

Characteristically, the schists are very well foliated mica-rich rocks which split easily along the foliation planes but tend to shatter when one tries to break them across the foliation. Most of the schists are very rich in muscovite, and contain varying amounts of plagioclase and biotite. They almost never contain potash feldspar but they commonly contain relatively abundant accessory minerals such as garnet, sillimanite, kyanite and staurolite. We have used the relative abundances of these accessory minerals

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[1] Uranium Resources Division, U. S. Geological Survey, Denver, Colorado.
[2] Earlham College, Richmond, Indiana.
[3] University of Wisconsin, Madison, Wisconsin.

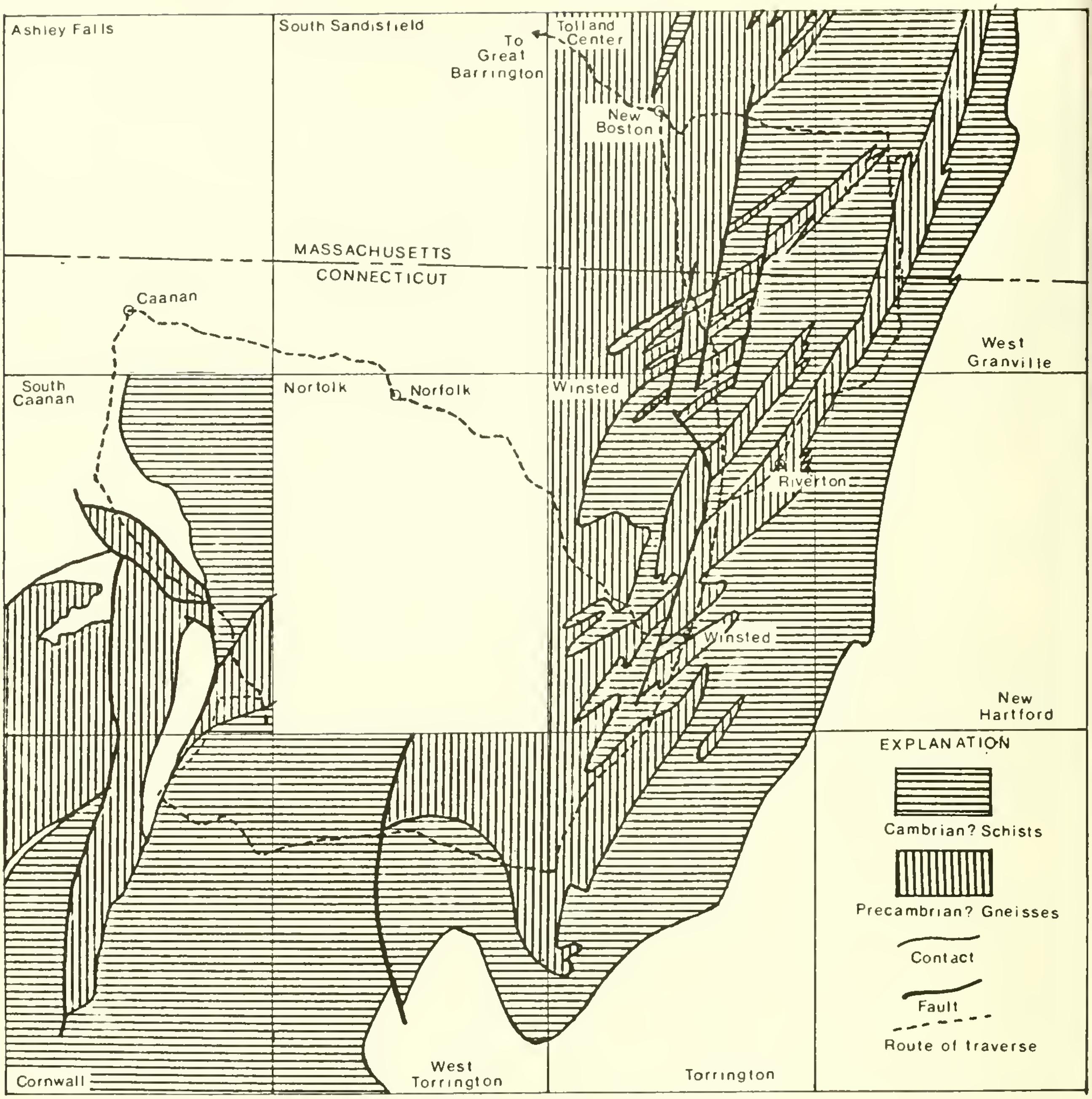

Figure 1: Generalized geologic map showing relations between Precambrian? gneisses and Cambrian? schists in western Connecticut and Massachusetts

between beds to evaluate structural and stratigraphic relations in parts of the area.

The only purpose of this trip, however, is to examine the contact relations between the schists and the gneisses. We believe the contacts that we will show you will convince you that they are normal stratigraphic contacts. We believe that in some places there may be faults associated with these contacts, but we also believe most of the evidence indicates that many of the units are in perfectly normal stratigraphic contact. We shall show our evidence and let you decide if we have enough.

If there were a direct method for establishing the ages of the units we would have no problem. If the gneisses are metavolcanics, at least in part, then perhaps they contain enough zircon for accurate dating, but we have no direct way to get at the age of the sediments. In addition, the rocks are all at sillimanite grade metamorphism, and one may well suspect that metamorphism has reset any radiometric clocks that may have existed in these rocks.

Road Log

This road log is designed both for those attending the field trip in conjunction with the NEIGC meeting of 1975, and to offer interested individuals opportunities to observe other localities where the phenomena herein described are well exposed, but inaccessible to large groups. Side excursions for those individuals who plan to make a more thorough independent investigation, are separated from descriptions of the main part of the trip by double asterisks (**). Those who opt to take this trip as individuals are advised to obtain the following topographic quadrangle maps: Tolland Center and West Granville, Connecticut and Massachusetts; and Winsted, New Hartford, Norfolk, South Canaan, Cornwall, and West Torrington, Connecticut.

Mileage

00.0 Center of New Boston, Massachusetts, at the intersection of State Routes 8 and 57, on the bridge over the West Branch of the Farmington River. (New Boston is about 30 miles east of Great Barrington and can be reached by going east out of Great Barrington on Route 23, and following Route 57.) It is assumed that you are eastbound on Route 57. Immediately after crossing the bridge the road makes a sharp right turn, and then you make a sharp left turn, following Route 57 and up the long hill climbing Tolland Mountain.

Although outcrops are sparse in the areas immediately adjacent to the road, most nearby exposures in this segment of the traverse are competent, hard, two feldspar gneisses.

00.6 Tolland-Sandisfield town line.

03.8 Tolland Center. **In the brush, about at the 1519 bench mark, southwest of the road intersection, is an outcrop of porphyroblastic garnet schist, containing garnets about ¼ to ½ inch in diameter. This schist is a relatively isolated occurrence of "Hoosac-type" schist well east of the main mass of Hoosac Schist. An additional side excursion

involves going north from Tolland Center on Clubhouse road, to Schoolhouse road, north on Schoolhouse road to Blandford road, east on Blandford road to Amos Case road, and left on Amos Case road to the 140 foot crest just west of Babb Hill. In the woods just to the west are abundant outcrops of both gneiss and schist, and many of these show contact relations between the rock types rather well. In addition, several different rock types commonly present in the Hoosac Schist, are abundantly exposed. A note of caution: Blandford road is not always passable, especially in the spring, and if a two-wheel drive vehicle is used care must be exercised before proceeding down Tow Hill.**

04.3 Intersection of Schoolhouse road, and Hartland road. **In the yard behind the school on the northeast corner of the intersection, are several small pavement outcrops that show very thin layers of gneiss and schist.**

04.4 Outcrops on both sides of the road are of Hoosac Schist.

06.6 Hubbard River, Granville-Tolland town line.**Outcrops on the east facing slope to the southwest, and in the stream valley to the north are poor, but some show the interlayering of the schists and gneisses. Contacts are not well exposed.**

07.0 West Hartland road, turn right (south) toward Granville State Forest.

07.5 Hubbard River. Turn left along road along north side of river. **In season there may be a parking fee, inasmuch as this is a State Park. Outcrops upstream along the river are both schists and gneisses, but in general they are poor and do not clearly expose contact relations.

08.0 Loop at end of road. At this time of year, parking should be no great problem, but the park road is narrow, and some caution should be exercised if we are to be able to turn around.

Outcrops are essentially non-existant along stream from bridge to this point. The area is interpreted to be underlain by Hoosac Schist.

Take trail from loop at end of road along east side of river going downstream. At about 2,000 feet, trail intersects high pressure gas line. Continue additional 500 feet along trail. In 1974, an abandoned boiler was alongside the trail at this point. Outcrops of interlayered schist and gneiss are on both stream banks from this point to about 200 feet downstream.

It is much slower to follow the stream valley, but there are nearly continuous outcrops of gray two feldspar gneiss from about the loop at the end of the road to the first schist outcrop.

After examining the outcrops, return to the cars and retrace route back to the main road at the bridge.

08.5 Main road, turn left (south).

10.8 State line. Marked by post on east side of road. Several outcrops in woods to west, of both gray gneiss and schist, but contacts are not well exposed.

11.9 Intersection with Connecticut State route 20. Turn right (south). Outcrops on both sides of road are of Hoosac schist with abundant pegmatite. We have seen no exposures of gneiss east of this road. According to our interpretation, we are near the top of the Hoosac.

12.7 West Street (Pinehurst Road on 1955 edition of West Granville quadrangle map). **At the south end, and in the woods to the north and east of Howells pond, which is near the end of West Street, are several rather poor outcrops of both schist and gneiss.**

13.3 New Hartford quadrangle boundary.

13.6 Junction with Connecticut State Route 181. Turn right (west), stay on Route 20.

14.7 Winsted quadrangle boundary.

16.4 Hogback road junction to right. Outcrops to left are of calc-silicate gneiss. These are included in the gneiss complex, they are not a typical Hoosac lithology.

17.0 Turn right (west) stay on Route 20. Hitchcock chair factory to right.

17.2 Riverton Connecticut. Turn left (stay on route 20).**About .1 miles farther, just across the bridge over the Still River an unnamed road to the left follows the West Branch on the Farmington River. The first outcrop, on the left side of this road, exposes an excellent contact between schist and gneiss. To the southeast along the road (mostly along the river bank) are scattered small outcrops of both schist and gneiss, but we saw no good contacts in this area. Continue to Pleasant Valley and turn right (south) on State Route 318. About 1.2 miles farther is junction of U.S. Route 44. Turn right (west). Outcrops on both sides of road are various gneisses. Continue west on Route 44 for about 2.4 miles. Turn left (south) on West Hill road. Along the west side of West Hill lake are scattered outcrops of Hoosac-like schist in a large area of gneiss. No good exposures of contacts were found in this area. Return to Route 44 and continue west to Winsted Connecticut. At the exit ramp of new Route 8 are several exposures of contacts between schist and gneiss, parking for more than one or two cars is difficult, however. Rejoin road log at junction of Routes 8 and U.S. 44 in the center of Winsted.**

19.3 Junction with State Route 8. Turn left (south).

21.2 Center of Winsted Connecticut. Junction with U.S. Route 44. Turn right (west).

23.1 Beginning of long road cut. Outcrops are mostly gray gneiss with a few thin layers of Hoosac-like schist near the east end, and abundant pink feldspar pegmatite and granite near the west end.

24.1 Entrance to Mad River Dam. Turn left through gate and park on top of dam. We will inspect the outcrops along the spillway where several exposures show the contact between schist and gneiss. When finished, return to cars and turn left (west) on Route 44. **If the gate across the road to the dam is locked, access can be gained by continuing west on Route 44 and making the first possible left turn. A secondary road goes east to the base of the dam, and access to the spillway can be gained by climbing the dam. Alternatively the outcrops along Route 44 can be studied, they show relations similar to those seen in the spillway cuts.**

38.1 Canaan, Connecticut, junction with U.S. Route 7. Turn left (south) on Route 7.

42.5 Junction with State Route 63. Bear left (south) on Route 63.

47.9 **Outcrops along road are of schist, on hills to east and south some contacts between schist and gneiss are exposed.**

49.8 Turn left on to dirt road and park where convenient. Walk to west (under powerline) to examine pavement exposures of contacts between gneiss and schist.

Return to cars and turn right (north) on Route 63.

51.2 Turn left on secondary road.

52.0 Junction with State Route 43. Turn left (south).

56.4 Junction with State Route 4, continue straight ahead (south) on Route 4.

56.6 Junction with Great Hollow road. Turn left (south).

57.1 Mohawk Mountain Ski area. Turn left into parking lot and park where convenient. Walk across the bridge, and then left (north) to northernmost ski trail. Walk up the ski trail and examine outcrops in area to north and east. In this area are many scattered outcrops of both schist and gneiss, contacts are not well exposed but area shows thinly layered nature of the schist-gneiss interbeds. Return to cars and turn right (north) out of ski area.

57.9 (Milage is approximate and depends on how far you drove in the parking lot). Junction with State Route 4. Turn right (north).

58.1 Junction State Routes 128, 43 and 4. Turn right (east) on Route 4.

70.9 Junction with Main Street, Torrington Connecticut. Road is marked as Route 8 on older maps. Turn left (north).

71.5 "Y" junction, bear right along old Route 8. Hills to northwest are underlain mostly by gneiss, with thin interbeds of schist. If there is time we will stop and examine these exposures.

80.5 Junction with U.S. Route 44, center of Winsted, Connecticut. Continue north across intersection. Join State Route 8.

92.1 Small, thin schist layers occur in the valleys between these big cuts. A few can be seen near the ends of the cuts. Most are rather poorly exposed.

92.1 Park as far off road as possible. Outcrop on west side of road is a coarse garnet schist in contact with gray gneiss. The schist is lithologically identical to schists in the Hoosac Schist to the east. After examining this exposure, return to cars, continue north on Route 8 to New Boston Massachusetts, and then west on Route 57 to Great Barrington, Massachusetts.

CROSS SECTION OF THE BERKSHIRE MASSIF AT 42^{o} N.: PROFILE OF A BASEMENT REACTIVATION ZONE[1]

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with Appendix presenting isotopic data by Douglas Mose[2]

Introduction

This trip will emphasize the structural geology and deformational history of Precambrian and Paleozoic rocks along a 30-km-wide traverse of the Berkshire massif at its widest point. The detailed stratigraphy of the Precambrian and Paleozoic rocks will not be treated here. These data were summarized in a previous N.E.I.G.C. trip (Ratcliffe, 1969) and can be found in the Stockbridge and Great Barrington quadrangle reports (Ratcliffe, 1974a, 1974b). Details of the blastomylonite and fold-thrust structures, as well as a general summary of the regional tectonics have been described by Ratcliffe and Harwood (1975). Bedrock mapping from 1967 to 1975 has been supported by grants from the Penrose fund of the Geological Society of America and by the U.S. Geological Survey in cooperation with the Massachusetts Department of Public Works.

The Berkshire massif, an area of Precambrian gneiss that extends from Adams, Massachusetts, south to Torrington, Connecticut, for a distance of about 100 km, forms one of the discontinuous exposures of basement rocks near the western limit of the intensely deformed and metamorphic core of the Appalachians from Newfoundland south to Alabama. This backbone of old rock in many places is at or near the contact between miogeosynclinal Cambrian and Ordovician shelf carbonate rocks and more eugoesynclinal rocks of the same age, and thus marks approximately the present locus of the Cambrian and Ordovician shelf edge at the eastern margin of the North American continent (Rodgers, 1968). At latitude $42^{o}15'$N. (figs. 1 and 2) the Berkshire massif is 25 km wide and tapers to the north and south where it apparently plunges beneath Paleozoic rocks (Herz, 1961; Gates and Christensen, 1965). Along the western contact, the gneisses overlie Lower Cambrian to Lower Ordovician shelf-sequence carbonate rocks of the Stockbridge Formation and the exogeosynclinal Walloomsac Formation (Middle Ordovician or younger) in low-angle thrust faults (Norton, 1969; Ratcliffe, 1969, 1974a, 1974b; Ratcliffe and Harwood, 1975). At the eastern contact of the massif, rocks of the Hoosac Formation (Lower? Cambrian or older) are thrust westward over the gneisses in a complexly interwoven fault zone (Trip B-4), although sedimentary contacts between gneiss and younger rocks may be preserved locally (Trip B-5). Thus, at this latitude, the mioeugeosynclinal transition and the shelf edge of Rodgers (1968) were originally located somewhere

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[1] Publication authorized by Director, U. S. Geological Survey.
[2] George Mason University.

above rocks that presently form the Berkshire massif, wherever they were located in Cambrian and Ordovician time. In addition, the depositional basin of the Taconic allochthon may also have been above gneissic rock now found within this zone (Zen, 1967). No data require this interpretation, and the depositional basin of the Taconic allochthon may well have been some unknown distance east of the restored position of the massif rocks.

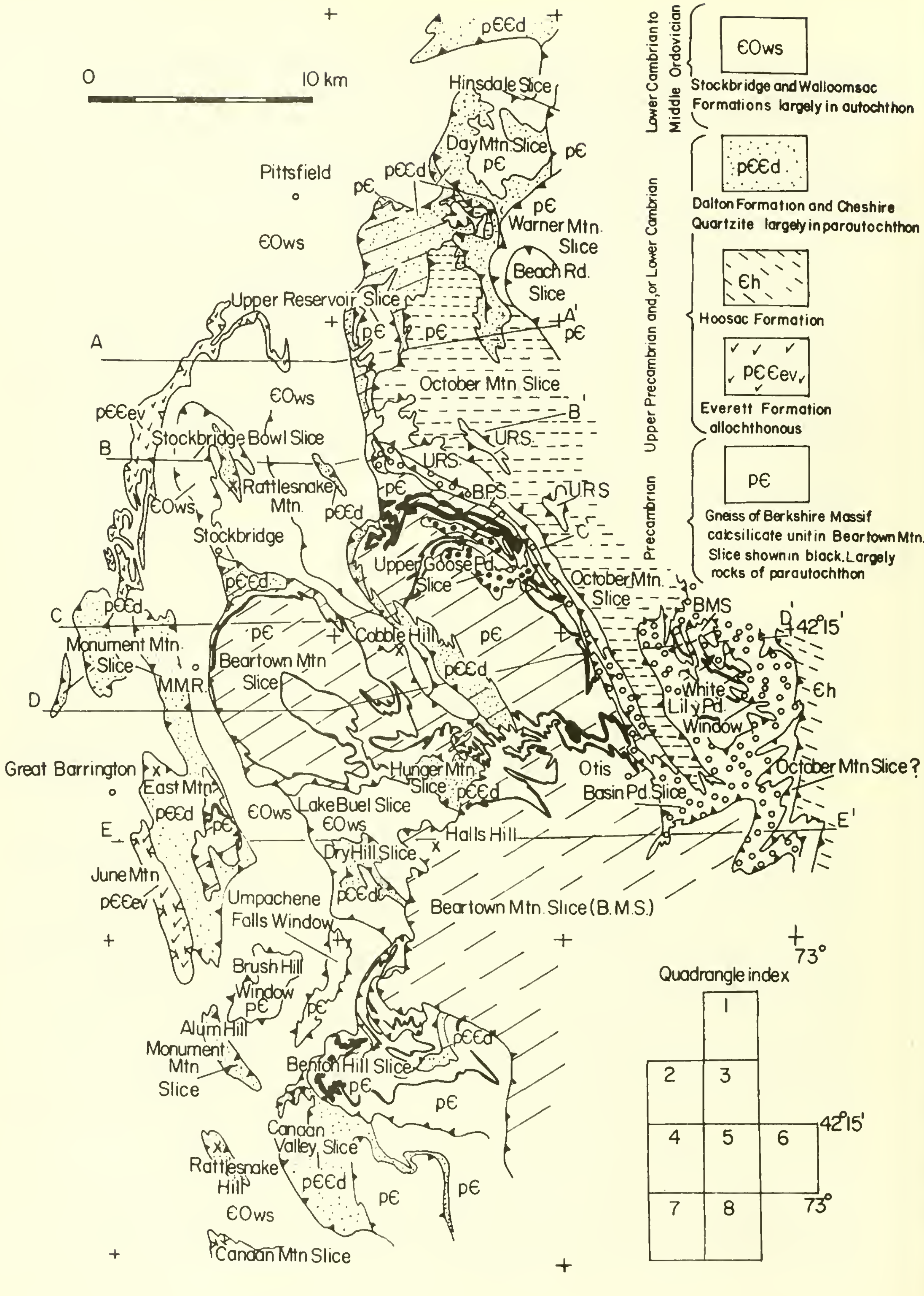
Solution of these major problems requires knowledge of the configuration of the continental margin in Early Cambrian to Middle Ordovician time. Palinspastic reconstruction requires knowledge of the amount and character of tectonic shortening, as well as tectonic bypassing preserved in the gneissic basement and cover rocks of the Berkshire massif before even qualitative reconstructions can be made. Some of the preliminary results of this study indicate that the basement rocks of the Berkshire massif record anomalously high amounts of Paleozoic strain in comparison with what is now known of the basement rocks of the Green Mountain anticlinorium to the north and Hudson Highlands to the south. The Berkshire Precambrian rocks have been remobilized in the Paleozoic (Ordovician?) (see Trip B-2) to produce a basement reactivation zone comparable in lateral extent and perhaps in magnitude of movement to the Pennine zone of the Swiss Alps and to that found in cores of collisional mountain belts in general.

Precambrian rocks of the Berkshire massif

Precambrian rocks of the Berkshire massif consist of a paragneiss sequence having interlayered mafic and felsic metavolcanic rocks. Distinctive and readily traced calc-silicate rocks, including thin zones of calcite and locally dolomitic marble, are found at several stratigraphic positions. This paragneiss sequence is intruded over a broad area by ferrohastingsite-biotite-microcline-quartz-oligoclase granitic gneiss, locally known as the Tyringham Gneiss (Emerson, 1899). The stratigraphic and intrusive relationships for rocks of the Beartown Mountain slice are shown diagrammatically in Figure 3.

Isotopic data for zircon from the Tyringham Gneiss and from a paragneiss unit, the Washington Gneiss (on Beartown Mountain), by R. E. Zartman of the U. S. Geological Survey are only slightly discordant and give Pb^{207}/Pb^{206} ages of 1040-1080 m.y. for both units (Ratcliffe and Zartman, 1971, in press). The close agreement in age between zircon from the intrusive rock and from the paragneiss suggests that (1) the Tyringham may have been intruded syntectonically within a period of dynamothermal metamorphism that reset the Washington zircon or (2) both the Washington and Tyringham zircons were reset by a postintrusive Precambrian dynamothermal event. The field data are consistent with either explanation. A recent Rb/Sr whole-rock isochron of about 1170 m.y. has been determined for the Tyringham Gneiss by Douglas Mose (Appendix 1) by using the Rb^{87} decay constant of 1.39 x 10^{-11} yr^{-1}. This suggests that (2) above may be the preferred interpretation of the zircon data. In addition, some effect on the U-Th-Pb systems during Taconic and Acadian metamorphism cannot be ruled out.

Metamorphic mineral assemblages and textures in the Precambrian rocks are generally consistent with the garnet-to-sillimanite grade of Paleozoic metamorphism. However, gneisses from the western edge of Beartown Mountain north to Pittsfield show retrogressive mineral textures. Locally, megacrysts



0
10 km
Pittsfield
ЄOws
pЄЄd
Hinsdale Slice
Day Mtn Slice
pЄ
Warner Mtn. Slice
Beach Rd. Slice
Upper Reservoir Slice
A
A'
October Mtn Slice
pЄЄev
Stockbridge Bowl Slice
B
B'
URS
BPS
Rattlesnake Mtn.
Stockbridge
Upper Goose Pd. Slice
October Mtn. Slice
BMS
C
C'
Monument Mtn. Slice
MMR
Beartown Mtn Slice
Cobble Hill
D
D'
42°15'
White Lily Pd. Window
Єh
Great Barrington
East Mtn
Hunger Mtn. Slice
Otis
Basin Pd. Slice
October Mtn Slice?
E
E'
Lake Buel Slice
Halls Hill
Dry Hill Slice
June Mtn
Umpachene Falls Window
Beartown Mtn. Slice (B.M.S.)
73°
Brush Hill Window
Alum Hill
Monument Mtn Slice
Benton Hill Slice
Canaan Valley Slice
Rattlesnake Hill
Canaan Mtn Slice
Lower Cambrian to Middle Ordovician
ЄOws
Stockbridge and Walloomsac Formations largely in autochthon
Upper Precambrian and, or Lower Cambrian
pЄЄd
Dalton Formation and Cheshire Quartzite largely in parautochthon
Єh
Hoosac Formation
pЄЄev
Everett Formation allochthonous
Precambrian
pЄ
Gneiss of Berkshire Massif calcsilicate unit in Beartown Mtn. Slice shown in black. Largely rocks of parautochthon
Quadrangle index
1
2
3
4
5
6
7
8
42°15'
73°

STRUCTURAL STACKING SEQUENCE OF THRUST SLICES IN THE WEST-CENTRAL PART OF THE BERKSHIRE MASSIF IN MASSACHUSETTS

At $42^{\circ}15'$

October Mountain Slice
(Washington and Tyringham Gneiss with unconformable Dalton-Cheshire sequence)

Upper Reservoir Slice (URS)
Tyringham Gneiss and unnamed hornblende gneiss and amphibolite unit

Basin Pond Slice (BPS)
Tyringham Gneiss with minor amounts of Washington Gneiss, and layered paragneiss

Upper Goose Pond Slice
Tyringham Gneiss

June Mountain Slice
Everett? Formation, structural position uncertain, perhaps uppermost slice locally involuted in Beartown Mountain slice

Beartown Mountain Slice (BMS)
Thick Precambrian sequence includes paragneiss, an important calc-silicate unit, Washington and Tyringham Gneisses, with unconformable Dalton-Cheshire sequence, forms a large nappe with westward vergence

Monument Mountain Slice
Dalton and Cheshire sequence with Stockbridge through unit b. Exposed as outliers from Rattlesnake Mt. in north to Rattlesnake Hill in south

Dry Hill and Hunger Mountain Slices
Precambrian gneiss, with unconformable Dalton-Cheshire sequence and Stockbridge through unit b. May be equivalent to Monument Mountain slice

Lake Buel and Stockbridge Bowl Slices
Stockbridge units a through e and Walloomsac Formation

Everett Slice of Taconic Allochthon

Rocks of the Stockbridge Valley
Dalton, Cheshire, Stockbridge, and Walloomsac. Locally at Umpachene Falls and Brush Hill windows detached from underlying Precambrian basement gneiss

At $42^{\circ}22'30''$ (see Trip B-9)

Beach Road Slice
Tyringham Gneiss with distinctive ferrohastingsite rodded aplitic facies, Washington Gneiss

Warner Mountain Slice
Washington Gneiss with unusually thick and persistent calc-silicate unit

Hinsdale Slice
Well layered paragneiss

Day Mountain Slice
Well layered paragneiss, Tyringham and Washington Gneisses. Unconformable Dalton with important coarse conglomerate, Cheshire Quartzite. Local interbeds of Hoosac-like quartz schists

October Mountain Slice
Washington Gneiss with poorly developed calc-silicate unit, and Tyringham Gneiss. Unconformable Dalton-Cheshire sequence

Dutch Hill Slice
Tyringham Gneiss with augen-rich blastomylonite at sole (may be equivalent to the Upper Reservoir slice)

Beartown Mountain Slice (BMS)
Largely Dalton-Cheshire sequence, with thick western facies of Cheshire Quartzite, and coarse basal conglomerate

Rocks of the Stockbridge Valley
Stockbridge and Walloomsac Formations no deeper tectonic levels exposed

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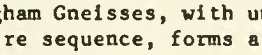
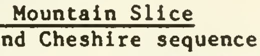
Fault symbols (teeth on higher plate, where faults are overturned teeth on original upper plate)



Premetamorphic (preM_1) thrust



Synmetamorphic (M_1) or older thrust with multiple movement history

Synmetamorphic (M_1) thrust formed during emplacement of Berkshire massif in Late Ordovician(?), thrust at base of Hoosac of preM_2 age

Index to numbered quadrangles and sources of data:

1. Pittsfield East (Ratcliffe, mapping in progress), 2. Stockbridge (Ratcliffe, 1974a), 3. East Lee (Ratcliffe, unpublished data), 4. Great Barrington (Ratcliffe, 1974b), 5. Monterey (Ratcliffe, 1975), 6. Otis (Ratcliffe, mapping in progress), 7. Ashley Falls (Ratcliffe and Burger, 1975), 8. South Sandisfield (Harwood, 1971, modified from Ratcliffe and Harwood, 1975).

Figure 1. Generalized geologic-tectonic map of a part of the Berkshire massif showing distribution of structural slices.

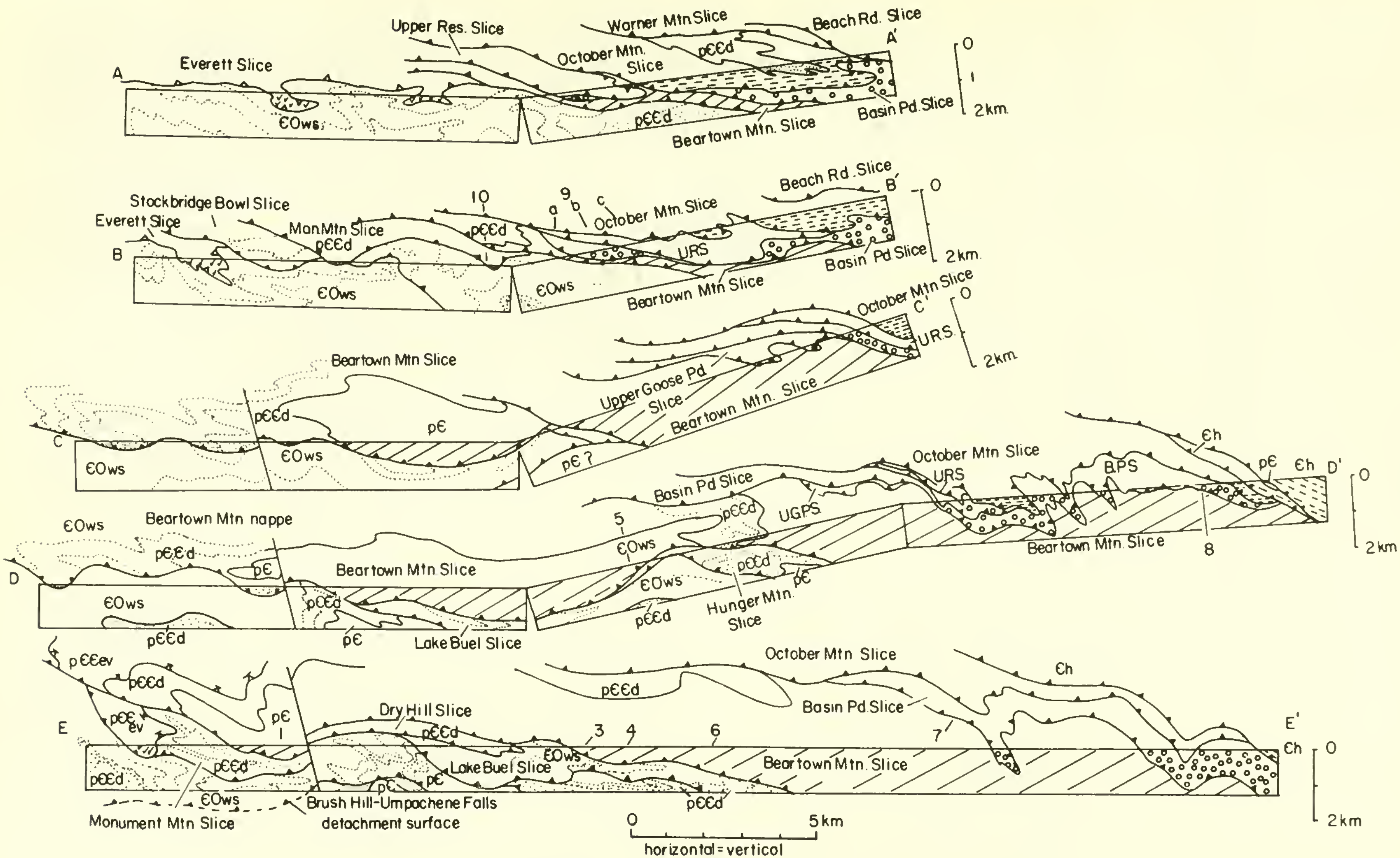


Figure 2. Cross sections of Berkshire massif to accompany geologic map, Figure 1. Fine dotted lines show form lines in Paleozoic rocks, symbols identified on Figure 1. Numbers indicate field trip stops, symbols for slices identified in explanation on Figure 1.

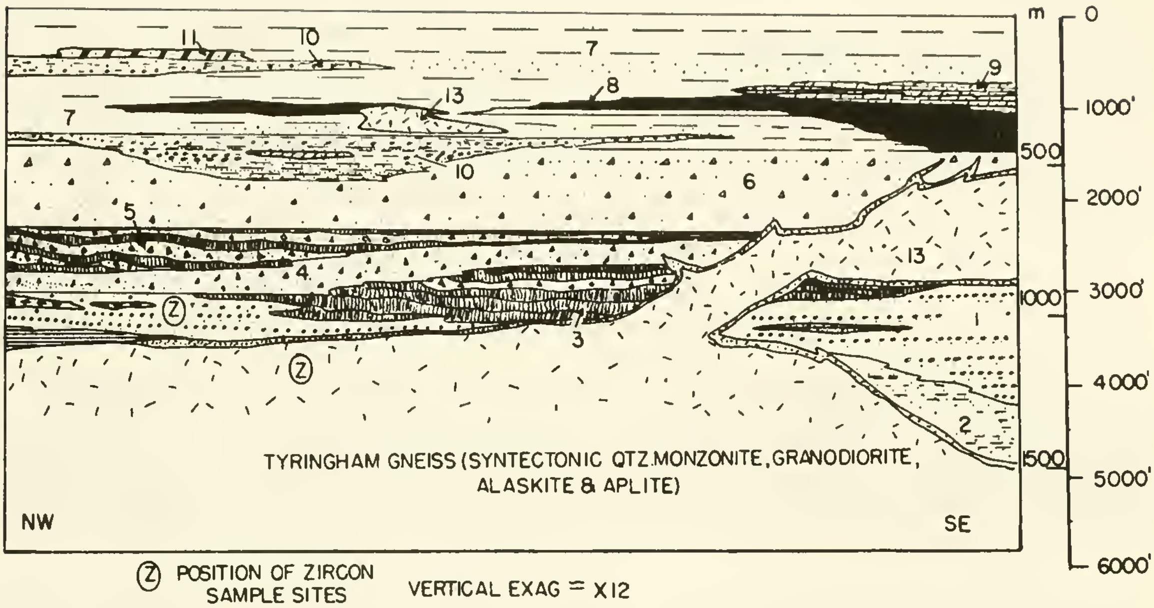


Figure 3. Composite stratigraphic section of the Precambrian rocks in the Stockbridge, Great Barrington, Monterey, and East Lee quadrangles, Massachusetts, showing inferred lateral relationships of lithologic units: 1 and 2, Washington Gneiss; 3, garnet-hornblende amphibolite; 4, leucocratic biotite gneiss and granulite; 5, garnet-hornblende-plagioclase amphibolite; 6, biotite-quartz-magnetite felsic gneiss; 7, biotite-quartz-plagioclase paragneiss (with interlayered lithologies 8 and 9); 8, calc-silicate rock and marble; 9, layered and spotted hornblende-plagioclase gneiss; 10, rusty-weathering schistose gneiss and granulite; 11, fine-grained garnet-hornblende amphibolite; 13, Tyringham Gneiss, intrusive biotite ferrohastingsite granodiorite with alaskitic chilled border facies.
Z shows position of zircon sample sites for U-Th-Pb study by R. Zartman (Ratcliffe and Zartman, 1971).

of perthitic alkali feldspar and sillimanite are preserved in the Paleozoic garnet and staurolite zones (Norton, 1969; Ratcliffe, 1969). This information, coupled with the observation that gneissic mineral textures are truncated by the unconformity at the base of the Dalton Formation of late Precambrian(?) and Early Cambrian age (Ratcliffe and Zartman, in press), indicate that the Precambrian rocks were crystallized and folded in a severe dynamothermal event in the Precambrian (for further details see Trip B-9).

Paleozoic stratigraphy is summarized in Table 1.

General structural geology of the Berkshire massif at 42° N. latitude

Three major lithotectonic sequences are recognized; allochthonous Taconic sequence (Everett Formation), autochthonous shelf and exogeosynclinal sequence (Dalton-Cheshire-Stockbridge-Walloomsac sequence), and the parautochthonous Berkshire massif and its attached metasedimentary cover rocks (largely Dalton Formation, Cheshire Quartzite, and limited exposures of Stockbridge units).

The threefold classification is less than satisfactory because nearly all the rocks are detached and, strictly speaking, are allochthonous. The terms "allochthonous," "parautochthonous," and "autochthonous" are therefore used in a general and relative way to signify rock sequences that differ in degree of tectonic displacement on the basis of either facies considerations and (or) geometric evidence of physical overlap. Although autochthonous shelf-sequence rocks are more or less in place, detachment from the basement rocks and intense imbrication by low-angle thrust faults are recognized.

Figure 1 shows the distribution, stacking order, and general stratigraphy of the major tectonic units. The "autochthonous" valley sequence, Stockbridge and Walloomsac Formations, crop out in the Stockbridge Valley west of the Berkshire massif from Pittsfield south to Canaan Mountain. Precambrian gneiss appears from beneath the Stockbridge in Umpachene Falls and Brush Hill windows. Here the Stockbridge is uncoupled from the basement along a detachment zone of unknown magnitude (Ratcliffe and Burger, 1975). These relationships suggest hat the autochthon may be entirely separated by low-angle faults from the true basement, and possibly no truly autochthonous rocks are present.

Many ancillary thrusts have formed in the "autochthon" along the western edge of the massif, and the area from Monument Mountain east to Halls Hill is dominated by low-angle overthrusts and by intense isoclinal and recumbent folding.

The Berkshire massif proper consists of a series of overlapping low-angle highly folded thrust slices, some of which contain an unconformable cover of upper Precambrian(?) and Lower Cambrian clastic rocks (Dalton Formation and Cheshire Quartzite). The lowest and most extensive slice of basement rocks, the Beartown Mountain slice, contains a distinctive stratigraphic sequence (fig. 3). Higher slices overlap one another to the north and east. Northeast of Otis, the October Mountain slice is downfolded in a major

Table 1. Brief summary of Paleozoic stratigraphy

Walloomsac Formation (Middle Ordovician or younger)

Exogeosynclinal sediments (included in ЄOws in fig. 1). Dark-gray to black, muscovite-biotite-plagioclase-quartz schist (Ow) and interbedded massive phlogopite-plagioclase microcline-calcite-quartz marble (Owm). Schist is sillimanitic in east. Unit unconformably overlies the Stockbridge Formation and bevels progressively deeper in eastern areas where local interbedded quartzites are found, indicating sediment source to the east.

Stockbridge Formation (carbonate shelf sequence) (Lower Cambrian to Lower Ordovician) (included in ЄOws in fig. 1)

Subdivided into lithostratigraphic units a through g. Units a, b, and c are largely dolomitic containing minor interbedded metaquartzites and schists; units d and f are distinctive crossbedded siliceous dolomite and calcite marbles that have abundant diopside in eastern areas; units e and g are largely calcite marble. Unit a is transitional through interbedding with the Cheshire Quartzite below.

Cheshire Quartzite (Lower Cambrian) (part of pЄЄd shown in fig. 1)

Massive vitreous quartzite that interfingers with feldspathic quartzites of Dalton Formation below.

Dalton Formation (part of pЄЄd shown in fig. 1) (Upper Precambrian? and Lower Cambrian)

Heterogeneous assemblage of massive feldspathic metaquartzite, quartzite, well-bedded tourmaline,muscovite-quartz flagstone, quartz-pebble metaconglomerate, vitreous metaquartzite, basal quartz pebble and arkosic metaconglomerate. Locally aluminous dark-colored garnet-muscovite-biotite-plagioclase-quartz schist similar to the Hoosac Formation is interbedded on East Mountain, at Stockbridge, and in the Canaan Valley slice where it is sillimanitic. The Dalton unconformably overlies gneiss of the Beartown Mountain and Benton Hill slices (Harwood, 1972), of October Mountain and Day Mountain slices, and locally occurs in the autochthon. The original depositional site of the Dalton in the Berkshire massif was well east of its present position.

northwest-trending synform. A complementary antiform in the north-central part of the Otis quadrangle exposes rocks of the Beartown Mountain slice in the White Lily Pond window. East of the White Lily Pond window, rocks of the Basin Pond slice dip gently east beneath an unnamed slice of Precambrian rock that may correlate with the October Mountain slice.

Windows at the east edge of the massif (fig. 2, sections D-D' and E-E') and the gently dipping folded configuration of the higher slices at this point suggest that the Beartown Mountain slice might have a similar geometry and be floored by a shallow thrust. If such is the case, the massif could be entirely allochthonous. However, no data are available at present that require this interpretation. Deep-core drilling to depths of 5 km in the White Lily Pond window might resolve this problem.

Rocks of the Hoosac Formation overlap several thrust slices in the gneiss at the east edge of the massif. Within 500 m of the gneiss contact, the Hoosac contains inclusions of gneiss as much as 100 m long. At the contact with these slivers Precambrian rocks and the Hoosac show an anomalous concentration of minor folds and a well-developed mullion structure which suggests the inclusions are tectonic. The contact beneath the Hoosac is interpreted as a gentle to moderately steep east-dipping thrust fault.

Direction of overthrusting

Slipline data have been determined, by using the techniques outlined by Hansen (1971), from many localities on the basis of the rotation sense of minor folds associated with the zone of blastomylonite, observation of mullions and slickensides, and study of lineations folded by similar folds. Some of the results pertinent to this trip are shown in Figures 4, 5, and 6. The azimuths of sliplines are fairly consistent, despite the effect of postthrust folding along NNW. and NNE. trends. These results suggest westward and southwestward thrusting generally consistent with previous assumptions based on the sense of overlapping (Ratcliffe and Harwood, 1975). The one slipline determination from the thrust zone at the base of the Hoosac is consistent with the movement pattern of the slices of the massif and suggests, but does not require, contemporaneous movement on all faults. The northernmost slipline measurement at the sole of the Hinsdale and Warner Mountain slices suggests that the longitudinal thrust component may increase northward toward the end of the massif. Further study will be necessary to resolve this point.

Chronology of structural and metamorphic events

Figure 7, reproduced from Ratcliffe and Harwood (1975), shows our conception of tectonic and metamorphic events that have affected this area.

Precambrian dynamothermal metamorphism about 1 b.y. ago, D_{pC}, postdated intrusion of the Tyringham Gneiss and produced the gneissosity in the Precambrian rocks. F_{pC} folds were generally east trending, having steep axial surfaces. Excellent exposures of the unconformity (seen on Trip B-9) at the base of the Dalton Formation of late Precambrian(?) and Early Cambrian age show a profound metamorphic and structural discontinuity.

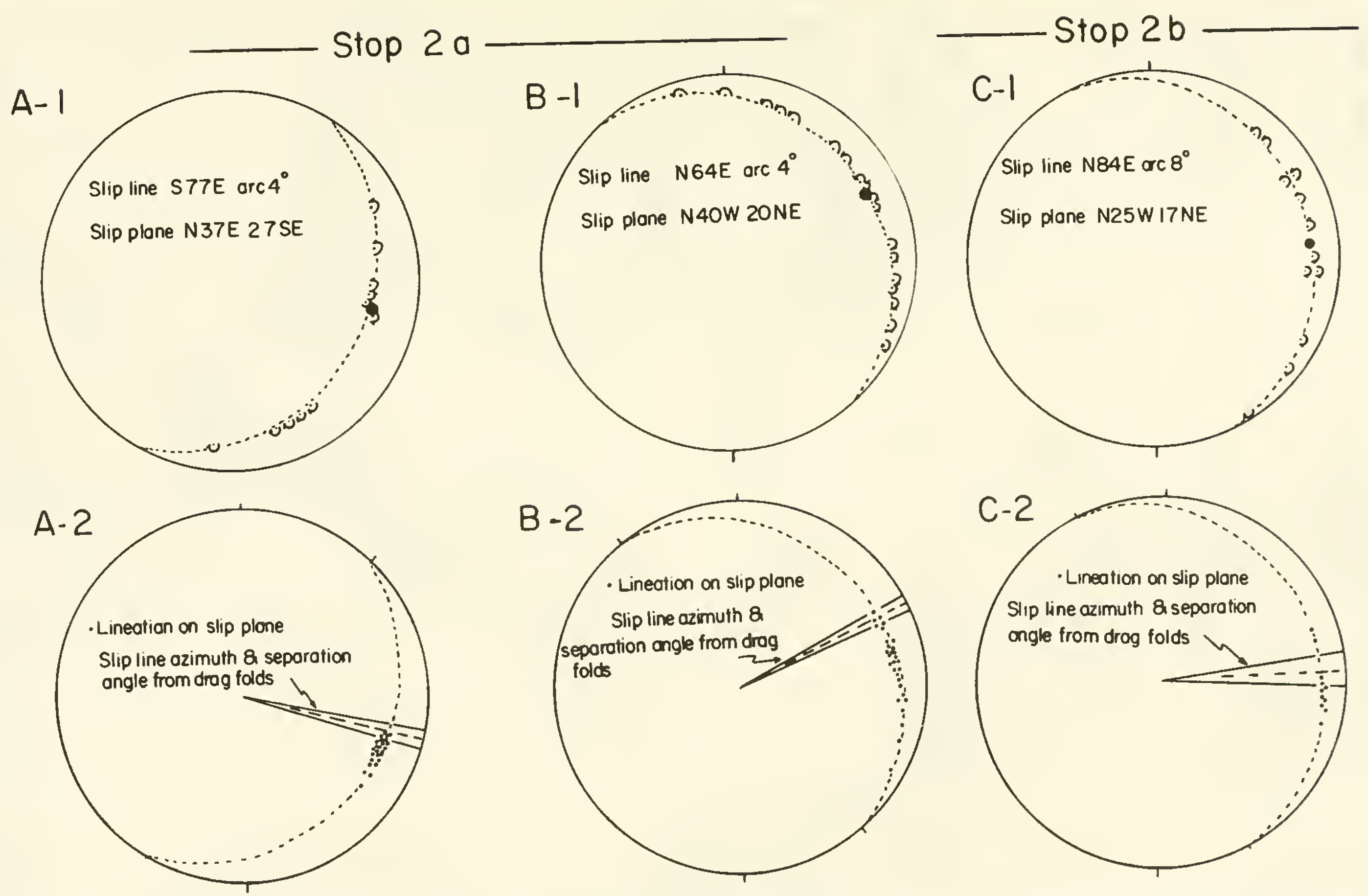


Figure 4. Lower hemisphere projections of plunge and rotation sense of F_3 minor folds at sole of Benton Hill slice at stops 2a and 2b. Lower diagrams show plunge of prominent mullion structures in blastomylonitic foliation and comparison with slipline and separation angle. Mullions approximate but do not agree with sliplines. Correspondence is greatest in most intensely deformed blastomylonite zones.

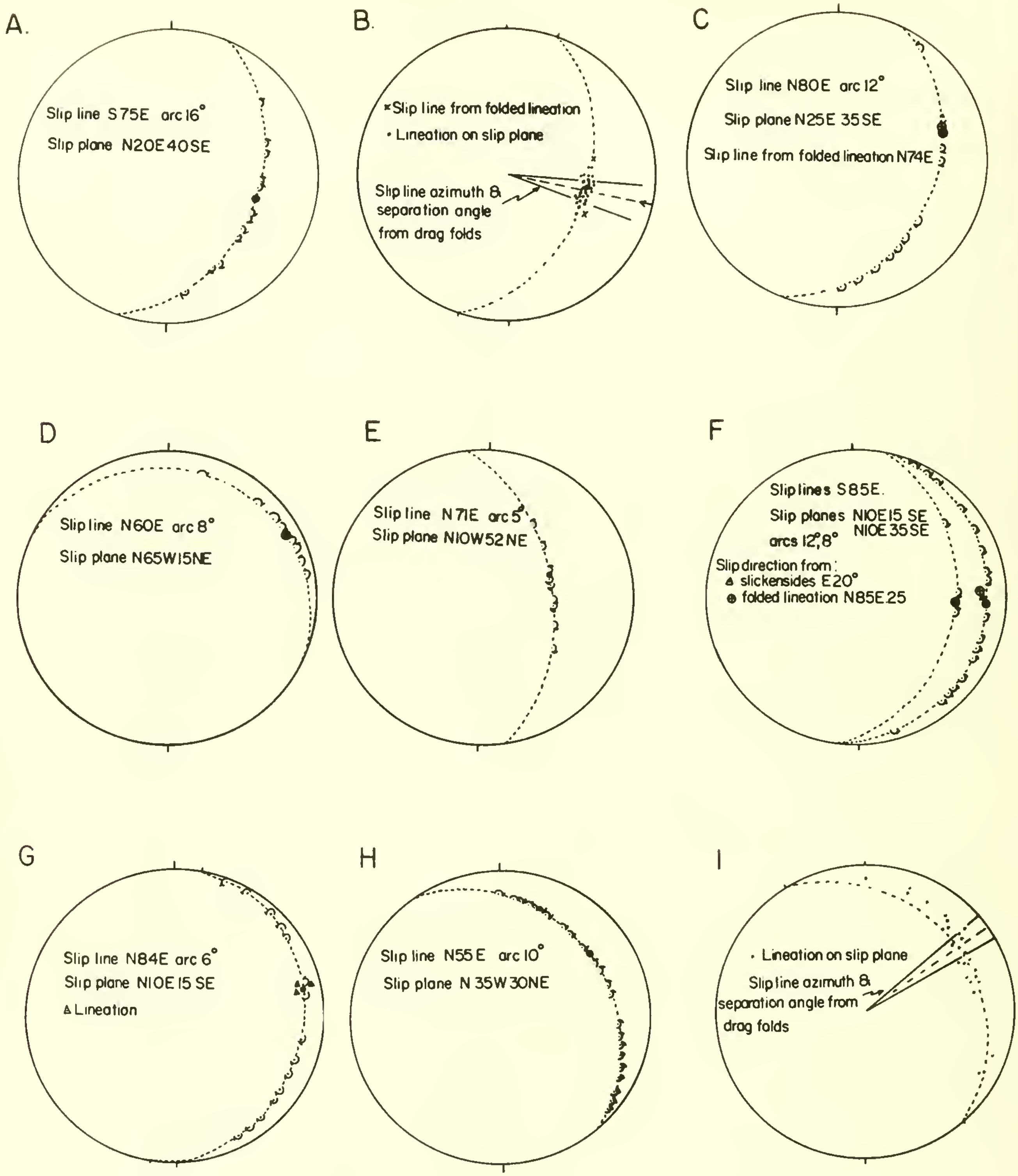


Figure 5. Slipline (solid circle) from rotation sense of F_3 folds, mullions or lineation dots, wear grooves, or slickensides identified on diagram. A and B (Stop 3), C (Stop 4, Halls Hill), D (Stop 6), E (Stop 7), F (Stop 9a), G (Stop 9c), H and I (Stop 10). Lower hemisphere.

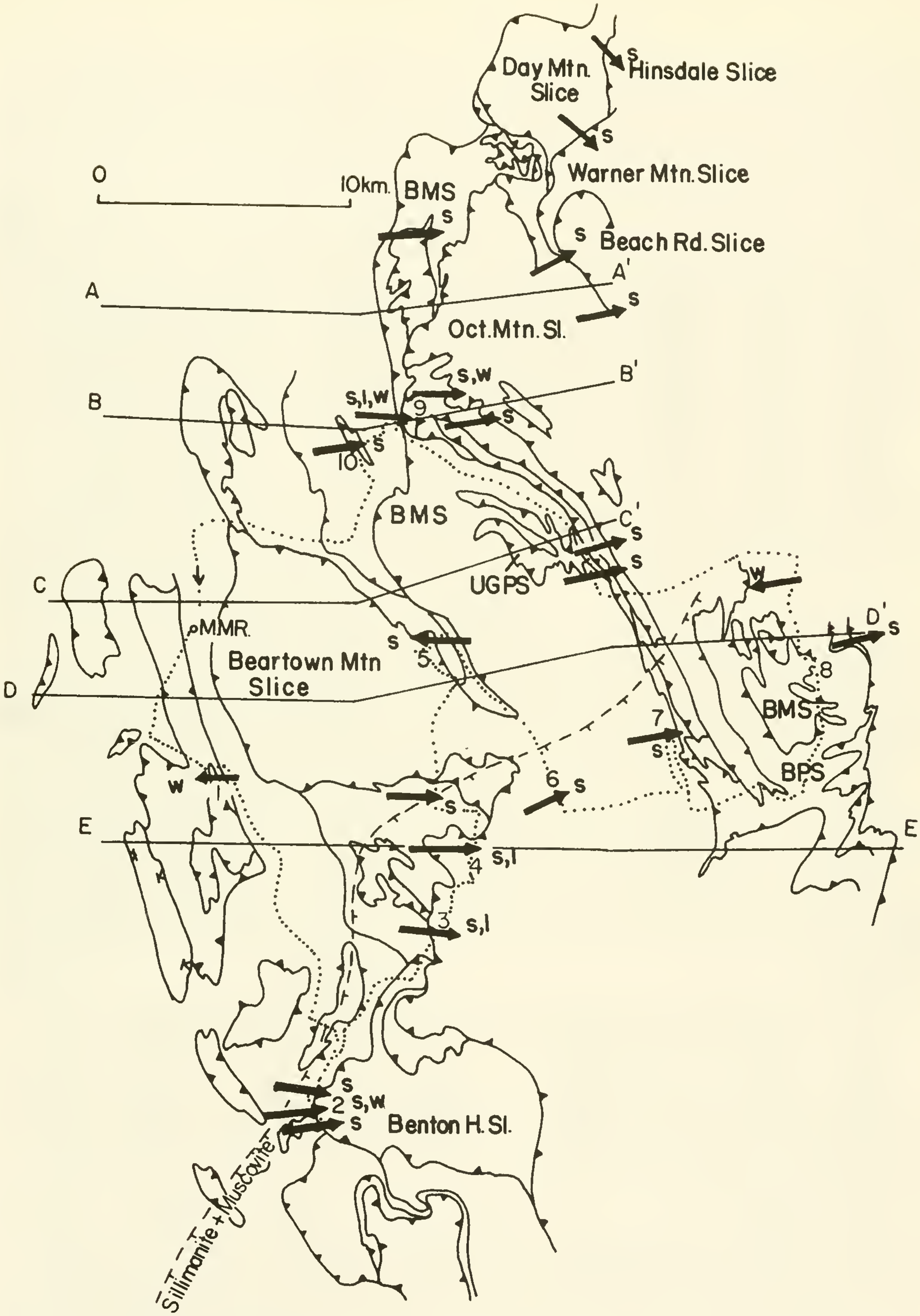


Figure 6. Generalized tectonic map showing thrust slices and slipline determinations: S (from separation angle of minor folds), W (from wear grooves), L (from folded lineation). Data for stops on this trip are presented in Figures 4 and 5. Stop locations and section lines of Figure 2 are identified. MMR--Monument Mountain Regional High School.

Deformational event	Number of fold system	Types of folds and areal extent	Important tectonic features	Metamorphic event	Important crystalloblastic and other structures	Igneous intrusion	Probable age of rocks in figure 1	Orogeny
D_6	F_6	North-south open folds of foliation locally recognized in Stockbridge valley	Northwest- and north-trending normal faults		Hematite-cemented breccias		Uncertain (Middle Devonian to Late Triassic)	
D_5	F_5	N 25°-40° E -trending upright to northwest overturned folds of foliation, with axial planar slip or crenulation cleavage. Folds recognized throughout area of figure 1 west to Mount Ida in SW corner of Kinderhook 15-minute quadrangle, N. Y., where Taconic unconformity is folded by N 40° E upright folds	Refolds thrust sheets and blastomylonitic foliation **Chatham fault**	M_2 Thermal maximum Acadian metamorphism	Crenulation of sillimanite alined in axial surface of F_4 folds, granulation of garnet and staurolite that includes F_4 foliation **mylonite of Taconic foliation**		Middle to Late Oevonian (Ratcliffe 1969a, b, 1972) **Ratcliffe, Bahrami (in press)**	Acadian orogeny
D_4	F_4	Northwest-trending upright to southwest-overturned folds with axial planar slip, crenulation, and flow cleavage. Folds recognized throughout area of figure 1, west to Chatham, N. Y., in center of Kinderhook 15-minute quadrangle	Folds thrust sheets and blastomylonitic foliation resulting in local overturning of thrusts, northwest-trending high-angle reverse faults		Muscovite, biotite realined and recrystallized in axial surface foliation, coarse sillimanite crystallized in foliation. Garnet, staurolite include folded F_2 fabric, and blastomylonitic foliation		Middle to Late Oevonian (Ratcliffe 1969a, b, 1972)	
			Granite crosscuts thrust fault and blastomylonitic foliation	Thermal maximum Taconic metamorphism	Granite lacks blastomylonitic foliation in country rocks	Granite stock, South Sandisfield quadrangle	Late Ordovician(?) (Harwood, 1972)	Phases of Taconic orogeny
D_3	F_3	Northwest-trending recumbent to strongly southwest-overturned folds of basement gneiss and large-scale southwestward thrusting of Precambrian rocks of Berkshire massif across autochthon. Fold and thrust style recognized from Windsor quadrangle, Massachusetts (Norton, 1969), south to Norfolk quadrangle, Connecticut (Harwood, unpub data), along west front of Berkshire massif	Faulted recumbent folds and nappes, mylonite gneiss, blastomylonite associated with major thrusts Thrust sheets at June and Canaan Mountains transported with Berkshire massif		Alaskite has weakly developed blastomylonitic foliation but intrudes more highly cataclastic rock in fault zones, mylonite gneiss, blastomylonite has muscovite, biotite, hornblende with lepidoblastic texture, cataclasis of F_2 foliation, thrusting synmetamorphic	Alaskite sills in faults and magnetite mineralization	Synchronous with latest movements or thrusts (Late Ordovician?) Thrusting probably late Ordovician based on age of cross-cutting granite	D
D_2	F_2	Isoclinal northeast-trending northwest-overturned to nearly recumbent folds with strong axial planar foliation which is dominant foliation in most autochthonous and allochthonous (Taconic) rocks, but not clearly present in Paleozoic rocks attached to Berkshire massif. Folds extend west to Mount Ida where unconformable beneath lowermost Oevonian	Folding of Taconic thrust contacts, regional foliation and refolding of slump or soft-rock folds in Taconic allochthonous rocks	M_1	Lepidoblastic muscovite, chlorite, biotite, and ilmenite in foliation, chloritoid, albite include foliation but are kinked by F_4 structures		Middle to Late Ordovician(?)	C Taconic orogeny
D_2(?)	F_2(?)	Folding and metamorphism of Lower Cambrian metasedimentary rocks attached to Berkshire massif and in independent thrust slices at June and Canaan Mountains	Coarse foliation or schistosity formed	M_1(?)	Muscovite, biotite lepidoblastic in schistosity		Time of metamorphism very uncertain depending upon original position of these rocks, and timing of tectonic events at that site (Middle Ordovician to Cambrian?)	(C)?
D_1	F_1	Intrafolial minor folds associated with Taconic thrust contacts. Soft rock or slump folds in Taconic allochthonous rocks, scale of pre-F_2 folds not determined but widespread, area shown in figure 1, west to Mount Ida	Emplacement of upper Taconic slices (here, Chatham and Everett slices) Emplacement of lower Taconic slices	No metamorphism recognized	Tectonic breccias with inclusions of Stockbridge Formation along thrusts (Zen and Ratcliffe, 1971) Wild-flysch-like sedimentary rocks along base of thrusts		Uncertain (Middle Ordovician?) Middle Ordovician (Zen, 1972b, table 1)	B A
		Middle Ordovician unconformity						
D_0		Warping of Lower Cambrian to Lower Ordovician carbonate shelf sequence; locally dips near vertical (Ratcliffe, 1969a); possible block faulting					Late Early to Middle Ordovician (Zen, 1972b table 1)	Pre-Taconic disturbance
		Pre-Dalton unconformity						
$D_{p€}$	$F_{p€}$	Isoclinal east-west-trending folds with generally steeply dipping axial surfaces and strong axial planar foliation, deformation of all Precambrian rocks including granitic intrusions such as Tyringham Gneiss	Gneissosity in Precambrian rocks of Berkshire massif	$M_{p€}$	Diopside sillimanite, hornblende, microcline, perthite formed in dynamothermal event	Granodiorite-quartz monzonite intrusions such as Tyringham Gneiss, syntectonic	Dynamothermal event and granite intrusion approximately 1.04 b.y. (Ratcliffe and Zartman, 1971)	Grenville orogeny
		Pre-Tyringham foliation						

Degree of basement participation and intensity of deformation increasing with time →

Figure 7. Chronology of tectonic events recognized in southwestern Massachusetts, northwestern Connecticut, and adjacent eastern New York (reproduced from Ratcliffe and Harwood, 1975).

Paleozoic deformations D_0 to D_5 include two distinct periods of dynamothermal metamorphism. The earlier one (M_1), Taconic orogeny phase C, was coincident with formation of the regional slaty cleavage and schistosity in all Paleozoic rocks and may have continued into phase D of the Taconic orogeny during the emplacement of the thrust slices of the Berkshire massif.

Minor alaskites and associated small magnetite deposits were generated at the time of thrusting (D_3) along the soles of the lower slices of basement rock. A granite stock in the South Sandisfield quadrangle (Harwood, 1972; Trip B-2) and preliminary zircon data by R. E. Zartman, U. S. Geological Survey, indicate that the granite may be Late Ordovician in age (Harwood, 1972). The age of the alaskites is less certain. Preliminary Rb/Sr whole-rock data from the alaskite zone at Stop 6 by Douglas Mose (see Appendix 1) are inconclusive but suggest the possibility of formation of the alaskite in the Ordovician. Further work will be necessary to resolve this problem.

Emplacement of the slices of the Berkshire massif was accompanied by intense recumbent folding of all rocks, but these thrusts and recumbent structures postdate earlier metamorphic (D_2) structures. Distinctive zones of intense recumbent folding (F_3) and cataclasis near thrust faults produces a distinctive fold-thrust fabric, accentuated by seams of blastomylonite (Ratcliffe and Harwood, 1975).

M_2, the second metamorphism, produced a Barrovian zonation from biotite to sillimanite grade, which postdates emplacement of the Berkshire massif because (1) isograds pass undeflected across the thrust faults, and (2) inclusion textures in garnet, staurolite, and biotite indicate that crystallization of (M_2) minerals was synchronous or later than F_4 cross folds that, in turn, deform the thrust slices.

Staurolite, kyanite, and, locally, sillimanite are found in Silurian and Lower Devonian rocks east of the Berkshire massif (Thompson and Norton, 1968; Hatch, 1972; Stanley, 1975, and Trip C-11; Norton, 1975, and Trip B-4), thus suggesting but not requiring that D_4 and D_5 events are Acadian. Available K-Ar and Rb-Sr mineral ages (Zen and Hartshorn, 1966) also suggest that M_2 may be Acadian.

Blastomylonitic textures, deformation structures, and igneous rocks associated with overthrusts

A distinctive fold-thrust fabric (Ratcliffe and Harwood, 1975), composed of vertically stacked isoclinal recumbent folds (F_3) having amplitude-to-wavelength ratios of as much as 20:1 and a penetrative blastomylonitic foliation are found in all rocks adjacent to the thrust faults at the west edge of the massif. A penetrative axial-plane foliation crosscuts older gneissosity and schistosity (in Paleozoic rocks), and locally fine-grained alaskites rich in magnetite intrude the fault zones. At many localities, the cataclasis and recrystallization disappears away from the thrust. Rocks in the fault zones are mylonite gneiss and blastomylonite following the terminology of Higgins (1971, p. 11-13). Excellent exposures of this fold-thrust fabric associated with different slices will be examined at many stops on this trip and on Trip B-2. The faults are folded, and, locally, the fold-thrust fabric and the faults are overturned.

Exposures of the fold-thrust fabric in the eastern part of the massif at the latitude of this trip contain abundant stringers of granite and larger bodies of foliated biotite-spotted quartz monzonite of uncertain age and origin (Stop 8).

Magnitude of overthrusts and regional implications

Cross sections (fig. 2) across the Berkshire massif near 42^{o} N. show my present structural interpretation. Sections are drawn approximately parallel to the determined sliplines. Sections D and E show that combined Beartown and Monument Mountain slices overlap the underlying autochthon for a distance of 21 km. In addition, the lower Lake Buel slice overlaps the autochthonous rocks for approximately 6 km, and the Dry Hill slice-Hunger Mountain slice overlap the Lake Buel slice by 5 km. This indicates a minimum shortening of 11 km in the "autochthon" beneath the Beartown Mountain slice. Detachment of the "autochthon" in the Brush Hill and Umpachene Falls windows requires additional shortening of unknown magnitude.

The 21-km movement for the Beartown Mountain slice serves as an estimate for minimum westward displacement of the Berkshire massif as a whole. The Dalton-Cheshire sequence and unit a of the Stockbridge are exposed well east of the leading edge of the Beartown Mountain slice, east of Tyringham (fig. 2, sec. D). The bank edge for Early Cambrian time must have been 21 km farther east than the present easternmost position of these rocks in the Beartown Mountain slice. Additional shortening of 7 km in the Paleozoic cover sequence attached to the Beartown Mountain slice indicates that the restored position actually is at least 28 km rather than 21 km to the east. This minimum position is shown in Figure 8. This reconstructed position lies approximately at the present position of the eastern edge of the Berkshire massif and above the crest of the regional Bouguer gravity anomaly.

Additional internal strain in the Berkshire massif, as recorded by the successively higher slices, is appreciable and may be used to estimate the original width of the Grenville basement now telescoped in the massif.

If we assume that the individual higher slices did not all root in the same zone, as the varying stratigraphy of the slices (fig. 2) suggests, the cumulative displacement may be summed to estimate the amount of shortening. The displacements calculated from the cross sections will yield minimum figures. However, estimates of lateral shortening must be less than the displacement if the original thrusts were not horizontal. Because of geologic uncertainties, the estimates obtained can only be regarded as approximate.

The Basin Pond slice overlaps and truncates structures in the Beartown Mountain slice for a distance of approximately 15 km, measured along its presently folded contact with the Beartown Mountain slice. Likewise, the October Mountain slice may overlap both Beartown Mountain and Basin Pond slices for an additional 16 km.

The part of the Beartown Mountain slice now exposed, after consideration

of shortening resulting from Paleozoic folding, represents an estimated width of approximately 30 km. Therefore, the original width of the basement rocks in the Berkshire massif at $42^{o}15'$N. could easily have been 60 km before Paleozoic thrust faulting. The entire massif must have been at a point at least 21 km east of its present position, thus placing the east edge 80 km east of Great Barrington.

A generalized palinspastic map of the Berkshire massif for the Early Cambrian, incorporating the estimates above (fig. 8), places the massif squarely above the crest of the regional Bouguer gravity anomaly (Kane and others, 1972). In addition, the eastern extent of the Dalton-Cheshire shelf facies, the shelf edge, and depositional basin for Taconic rocks are shown to lie at or east of this anomaly. In areas such as the northern end of the Green Mountain-Sutton Mountain anticlinorium and in the Manhattan Prong, where the basement rocks are regarded as essentially autochthonous, the position of the bank edge proposed by Rodgers (1968) coincides with the positive gravity anomaly. This suggests the possibility that the gravity anomaly, believed to be of deep crustal origin (Diment, 1968), may represent the long-preserved results of extensional necking of sialic crust and concomitant intrusion or upward flow of dense subcrustal rocks in the initial breakup of the North American continent in the latest Precambrian.

Although volcanic rocks are not common in Cambrian rocks east of the massif at this latitude, metabasalts are sparingly present in the Hoosac-Rowe sequence and in the Nassau Formation of the Taconic allochthon. The coarse graywackes of Rensselaer associated with Taconic basalts were derived from a western source area of Grenvillelike gneiss and could well have been deposited either above rocks now making up the higher slices of the Berkshire massif or above the eastern edge of the Beartown Mountain slice. Thus, what Cambrian volcanic rocks are present at this latitude could originally have formed at or near the present position of the positive Bouguer anomaly.

Alternately, this anomaly could represent the root zone of the Berkshire massif, in which deep crustal rocks, initially emplaced during drifting, were brought nearer the surface by a combination of high-angle faulting and warping during continental and (or) island-arc collision in the Taconic orogeny.

Origin of the anomalously high strain in basement rocks of the Berkshire massif

In intensity of basement reactivation, grade of dynamothermal metamorphism, and prevalence of recumbent folds, the Berkshire massif differs markedly from the Green Mountain anticlinorium to the north and from the northern end of the Reading Prong to the south. All these areas were deformed in the Ordovician. Basement reactivation in the Berkshires also is an Ordovician rather than Devonian event (Harwood, 1972; Ratcliffe and Harwood, 1975).

An uneven pattern of strain recorded in the basement rocks of western

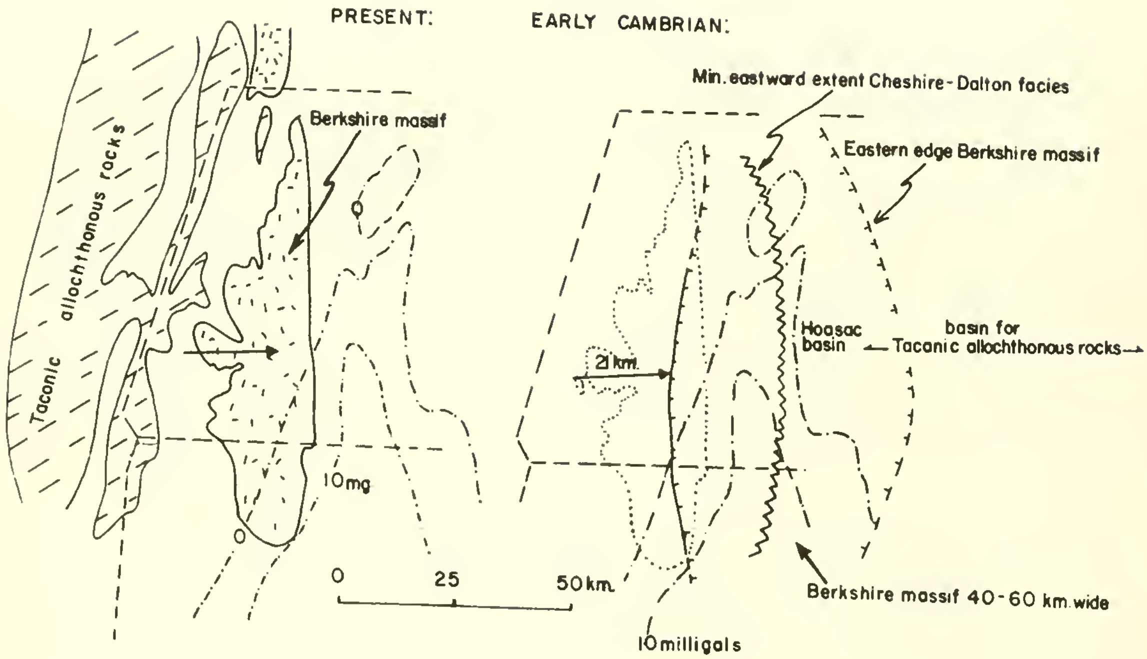


Figure 8. Palinspastic map showing approximate position in Early Cambrian time for Precambrian rocks of the Berkshire massif, minimum eastward extent of Dalton-Cheshire shelf facies, and possible position of depositional basin of Taconic allochthonous rocks with respect to the crest of the positive regional Bouguer gravity anomaly. Dot-dash lines represent 0 and 10 milligal gravity contours.

New England and New York suggests collisional effects of one or perhaps two plates having irregular margins after the general mechanism outlined by Dewey and Burke (1974). Analysis of this strain and reconstruction of the Cambrian shelf facies suggest that the continental edge of Cambrian and Ordovician North America may have extended oceanward in a promontory at the present latitude of the Berkshire massif.

The model proposed for the evolution of the Berkshire massif undoubtedly has imperfections, but the basic framework is based on detailed field data. The interpretations regarding net movement and geometric complexities are intentionally conservative and suggest approximate minimum constraints. The level of strain seen in the Berkshire massif seems to require continental and (or) arc collisional tectonics presumably in the Ordovician.

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Road Log

Mileage

0.0 Start log in parking lot Monument Mountain Regional High School, Rt. 7, Great Barrington. Turn left on Rt. 7.

3.8 Turn left (east) on Rt. 23 at yellow blinking light.

6.0 Park at second large roadcut on Rt. 23, just past entrance to Butternut Basin.

Stop 1. Outlier of Precambrian gneiss (pCbg, fig. 6) brow of Beartown Mountain nappe.

These exposures mark the westernmost extent of the core rocks of the Beartown Mountain nappe that here are detached from the Dalton and are thrust out over the Paleozoic rocks. Here the thrust surface is folded and dips northwest.

A large F_4 antiform plunges northwest, having an axial plane of N. 45° W., 75° NE. The dominant feature in outcrop is the folded gneissic layering, a relict of Precambrian dynamothermal metamorphism. However careful examination reveals that a relatively strong cataclastic and metamorphic (blastomylonitic) foliation outlined by several zones of bull quartz actually crosscuts the older structure to form cataclastic seams. Both are folded by the F_4 antiform and the blastomylonitic foliations dip SW., NW., and locally NE. Minor isoclinal folds and the blastomylonite are related to small dislocations that probably are subparallel to the major fault (perhaps 200 feet) beneath this outcrop.

Examination of the gneiss near the quartz pods reveals two kinds of lineation on the micaceous blastomylonitic surfaces: (1) an intersection lineation (gneissic layering x blastomylonite) that commonly plunges NNW. and (2) a wear-groove and mineral lineation that trends roughly E. The second lineation is consistent with the regionally determined slipline determinations (fig. 6) and probably records movement direction on the Beartown Mountain nappe at this point. Both lineations are folded in the antiform and locally are bent in minor folds that crenulate the blastomylonite.

This outcrop illustrates the relatively subtle thrust fabrics found in many rocks, and these features are easily overlooked. The late folding in the northwest direction produces axial-planar slip cleavage in the pelitic rock; biotite, garnet, chloritoid, and staurolite locally show inclusion textures, indicating that crystallization outlasted the crenulation. This late refolding probably is Acadian (M_2 of fig. 7).

As we walk southeastward up the plunge of late folds, the underlying Dalton appears from beneath the thrust. Within approximately 100 feet of the Dalton, the gneiss and Dalton are progressively more cataclastically deformed and difficult to tell apart. Excellent exposures of silvery-gray tourmaline-rich muscovite-biotite-plagioclase-quartz blastomylonitic schist (probably sheared Dalton) having abundant isoclinal F_3 "foldthrust" folds can be seen in the outcrop to the east on the south side of the road. Again, however, a late F_4 antiform has folded the blastomylonite and the axial surfaces of the F_3 folds. This rock is illustrated in Ratcliffe and Harwood (1975, fig. 6). Highly deformed, more normal Dalton underlies the gneiss 200 feet south of the highway behind the houses.

6.3 Proceed east on Rt. 23 0.3 mi. to Lake Buel Rd. Turn right. In 3.2 miles, at triangle, turn right onto Mill River Rd. Turn is poorly marked. Follow Mill River Rd. 1.9 mi. south and bear left at next

triangle (avoid turn to Sheffield Rd.). Stay on Mill River Rd.

12.8 Turn right at Mill River onto Clayton Rd. at "T" intersection. Cross bridge immediately, bear left (south) on Clayton Rd.

14.2 Turn left onto dirt road (first intersection south of Mill River; red house and barn on right side of road). Cross Konkapot River, past Umpachene Falls. The excellent exposures of sheared Precambrian gneiss and overlying Dalton-Cheshire-Stockbridge sequence were described previously (Ratcliffe, 1969, Stop 10). However, the "pebble conglomerate" at the base of the Cheshire is now identified as porphyroclastic mylonite gneiss, and the contact here is not a normal stratigraphic one, as I believed earlier. Although the geology is locally complicated by thrust faults of gneiss over Paleozoic rocks, I now believe that the Paleozoic rocks are detached from the basement by a major fault. In this area we are seeing the lowest tectonic level exposed in the Berkshires. Rocks of the Benton Hill area (Beartown Mountain slice), Stop 2, project over our heads.

14.8 At bridge turn right, bear right heading south onto Canaan-Southfield Rd.
15.8 Road branches in 1 mile by red barn; bear left.
17.3 At first intersection turn left onto Cross Rd. to Canaan Valley.

17.9 0.6 mi., pull into circle on left of road by cottages. Stop 2, Benton Hill--Torano property. Please obtain permission if you plan to return.

Stop 2. Benton Hill. Imbricate slices beneath Benton Hill slice, blastomylonite and fold thrust (F_3) fabric. See inset (fig. 9) for traverse and location of substops.

This traverse begins in OCsa of the Stockbridge. The carbonate rocks here actually are detached by a major thrust from the basement rocks that appear in the Umpachene Falls and Brush Hill windows. Above the dolomite marble are exposures of the Walloomsac (Owm-basal, feldspathic micaceous marble unit, and Ow-sillimanite-biotite-plagioclase-quartz schist). The Walloomsac here forms a parautochthonous sliver above the slice of Stockbridge but below three additional slices at the sole of the Beartown Mountain slice: (1) a small sliver of amphibolite gneiss, (2) a slice of Dalton, and (3) the uppermost Beartown Mountain slice that consists of a sequence of layered gneisses. The upper slice truncates all three of the lower slices and rests, in turn, on Stockbridge, Walloomsac, amphibolitic gneiss, and Dalton. Throughout the entire stack, an intense late east-dipping fold-thrust foliation has formed as the axial plane of (F_3) folds of schistosity and gneissosity. Near the base of the upper slice, intensely developed zones of blastomylonite can be seen parallel to the exposed thrust.

2a. Exposures of contact of Beartown Mountain slice on Walloomsac (Owm). In the small cliffs above are highly deformed gneiss having abundant F_3 folds and black blastomylonitic seams. Slipline determination using the rotation sense of minor folds and a plot of the prominant mullion structure are shown in Figure 4, A-1, A-2. To the

north at the sinkhole and collapsed cliff the contact of the gneiss and the underlying Owm is exposed in an upright late F_5 antiform. The contact is marked by a whitish zone of diopside, and tremolite-actinolite-calc-silicate about 8 cm thick. At the actual contact, no shearing is recorded; instead, the contact is healed by postthrust recrystallization (possibly a metamorphosed gouge). Benton Hill is east of the (M_2?) sillimanite isograd, which postdates the thrusting. Thus, we are looking at the effect of postthrust folding and metamorphism. Mullions are folded by the F_5 fold. Slipline determination and mullions from the cliff above the collapsed zone are shown in Figure 4, B-1 and B-2.

2b. Walk south along cliffs to excellent exposures of banded blastomylonite and, locally, quartz-potash feldspar segregations in blastomylonite. Surfaces of the quartz pods have wear grooves plunging S. 88° E. at 20° consistent with the slipline determination (fig. 4, C-1, C-2). Various kinds of blastomylonite showing various degrees of recrystallization can be seen.

Walk west back down hill toward cars.

2c. Amphibolitic gneiss of lowest sliver resting on Walloomsac having highly deformed schistosity. At base of the gneiss is "pebbly" looking biotite gneiss that actually is porphyroclastic blastomylonite. Fifty feet to the south a sliver of highly folded quartzitic Dalton (pCCdq) rests on the amphibolite and shows excellent F_3 folds and intense shearing. The Dalton is overlain by gneisses of slice 3.

18.5 Return on Cross Rd. to Canaan-Southfield Rd. and turn right.

19.5 Turn left and cross bridge onto Hadsell Rd. Follow Hadsell Rd. north 0.9 mile to intersection with two paved roads.

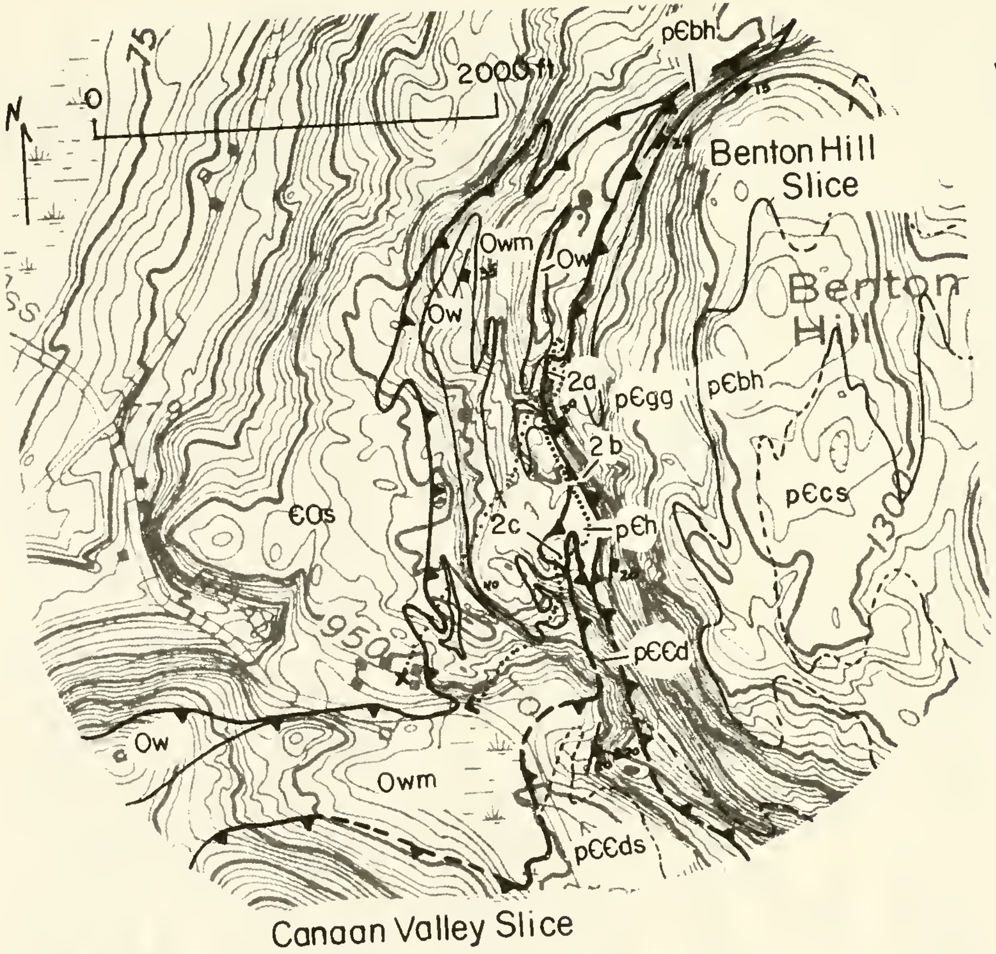
20.4 Turn right onto second paved road (northernmost), the Southfield-Mill River Rd.

21.6 In 1.2 miles Y intersection and stop sign. Turn left on road to New Marlborough.

23.0 Town of New Marlborough. Stop sign intersection with Rt. 57. Continue straight on dirt road headed north, New Marlborough-Monterey Rd. Slow down near crest of hill. Descend to flat and park alongside dirt road. Exposures are in small cliffs east of road.

25.1 Stop 3. Optional stop (parking here presents real problems). Sheared Washington Gneiss sole of Beartown Mountain slice. Ledges above road are rusty biotite-muscovite-plagioclase-quartz gneiss of the Washington Gneiss. An intense blastomylonite foliation and F_3 folds have formed. Locally amphibolite marker beds show the intense deformation. Slipline determination from rotation sense of F_3 minor fold is shown in Figure 5, A,B. To the west are rocks of the Dry Hill and Lake Buel slices.

26.1 Intersection at base of hill. Continue straight on narrower dirt road.



Ow, Owm Walloomsac Formation (Sillimanite schist and phlogopite marble)	ORDOVICIAN
COs Stockbridge Formation	CAMBRIAN
pCCd, pCCds Dalton Formation (metaquartzite and schist)	CAMBRIAN AND PRECAMBRIAN(?)
pCgg Granite gneiss pCbh Biotite-hornblende gneiss pCcs Calc-silicate rock and marble pCh Hornblende gneiss	PRECAMBRIAN

Thrust fault, teeth on upper plate

20
20

Strike and dip of axial plane of fold-thrust fabric, plunge of axis of reclined F_3 fold.

Figure 9. Generalized geologic map of the Benton Hill area, Stop 2. Dotted line shows traverse.

26.9 Intersection with paved road (Sandisfield Rd.). Turn left, 0.1 mi. Park just past house on left.

27.0 Stop 4. Halls Hill. Exposures of Tyringham-like granitic gneiss and fold-thrust fabric.

A small F_3 recumbent and reclined fold of gneissic layering is exposed in roadcut. In the axial part of the fold, a strong secondary metamorphic fabric has new oriented biotite parallel to zones of blastomylonite. The mineral textures in this rock, pictured in Figure 13 of Ratcliffe and Harwood (1975), indicate recrystallization of new brown biotite, apatite, clinozoisite, granulated quartz, and magnetite in the seams of blastomylonite. Superb F_3 folds can be seen in the cliffs in the woods to the east of the road. Locally sheared-off limbs of F_3 isoclines contain fine-grained "alaskitic" seams having parallel borders of black blastomylonite. Structures like this were seen at Benton Hill and will be seen at Cobble Hill (Stop 5) and at Stop 9. Evidently, the process that produces the "alaskites" operates over distances of centimeters to tens of meters. Slipline determinations from the cliffs northeast of the road are shown in Figure 5C. Analysis of folded lineations provides an independent determination of the slipline in agreement with that determined from the separation angle.

Continue north on Sandisfield Rd.

28.3 Intersection Rt. 23. Turn left.

28.5 Town of Monterey. Turn right at church toward Tyringham.

29.1 Branching road to left. Continue straight.

32.0 At the base of long downgrade (prepare to turn). Turn left on McCarthy Rd., dirt road.

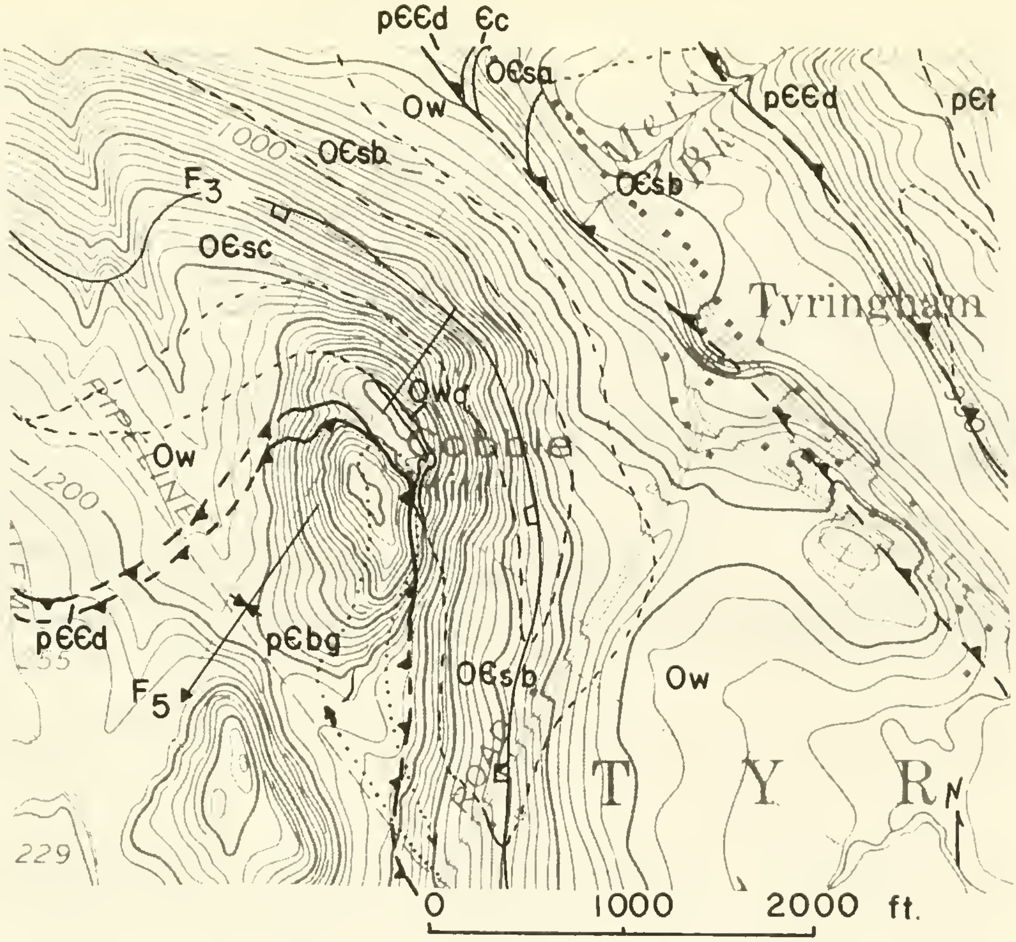
33.3 At triangle before farm turn right.

33.6 Park at gas-transmission-line crossing.

Stop 5. (Lunch stop.) Cobble Hill. Trailing edge of the Beartown Mountain slice in contact with Walloomsac of the autochthon and blastomylonite-alaskite zone.

Biotite-hornblende-granitic gneiss and biotite-quartz-plagioclase paragneiss of the Beartown Mountain slice overlie highly sheared rocks of the Walloomsac Formation and of the Dalton. Carbonate rocks of the Stockbridge are exposed at the base of the cliffs to the north and east. The thrust slices have been folded by a late F_5 synform.

At the north-facing cliff, rocks immediately above the thrust show excellent blastomylonite, F_3 folds, and a remarkable zone of alaskite and magnetite mineralization similar to the minor structures seen at Stop 4. Figure 10 shows a sectional view of this exposure. The alaskite forms a central zone between two symmetrical zones of



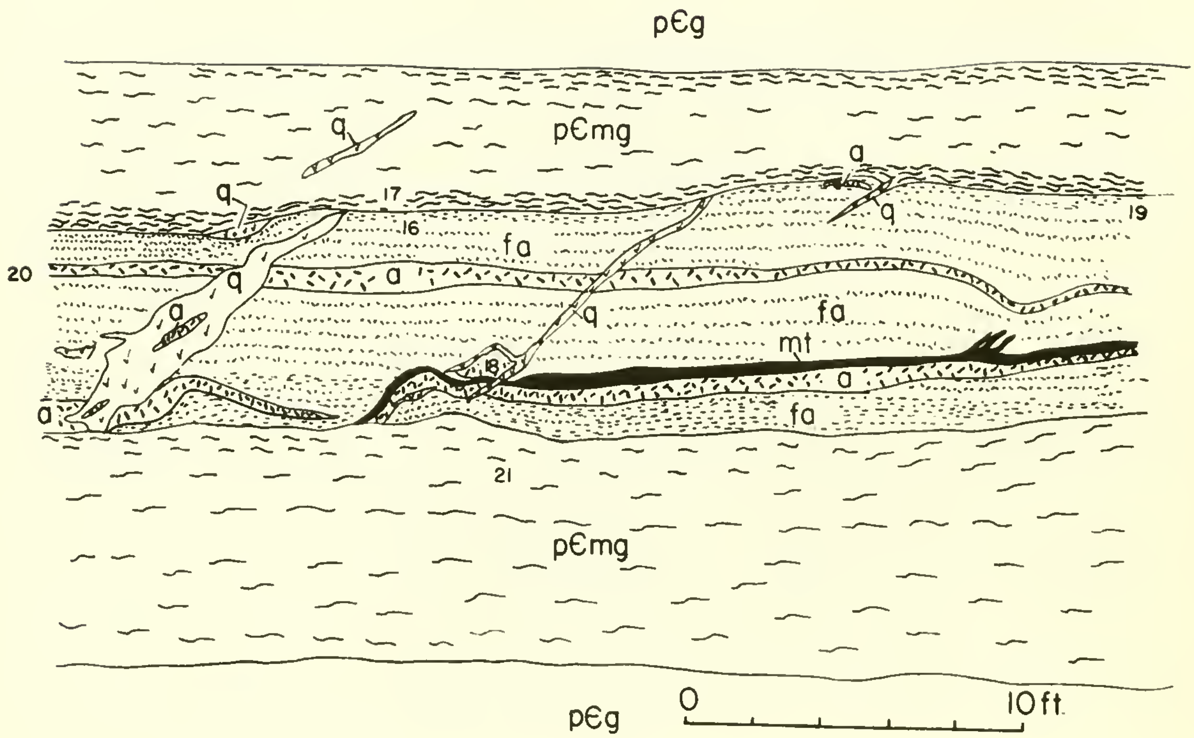
Ow
Owq
Walloomsac Formation
schist and quartzite — ORDOVICIAN
unconformity
OCsc
OCsb
OCsa
Stockbridge Formation — CAMBRIAN
Cc
Cheshire Quartzite — CAMBRIAN
pCCd
Dalton Formation — CAMBRIAN AND PRECAMBRIAN(?)
pCt
Tyringham Gneiss
pCbg
Biotite quartz plagioclase
paragneiss — PRECAMBRIAN

Axial trace F_3
recumbent fold

F_5 synform

Thrust fault,
teeth on upper plate

Figure 10. Sketch map of Cobble traverse, Stop 5. Dotted line shows traverse.



pЄg - biotite-microcline-plagioclase-quartz gneiss containing layers of hornblende gneiss, rocks show relict Precambrian gneissosity

pЄmg - dark-gray mylonite gneiss and hornblende gneiss containing abundant seams of fine-grained blastomylonite; more closely spaced symbols show zones of intense shearing

fa - light-gray to white, finely laminated and weakly foliated, biotite-muscovite-microcline-oligoclase-quartz "alaskite" containing scattered magnetite; more closely spaced symbols show better foliated rock.

a - light-gray, massive, poorly foliated, biotite-muscovite-oligoclase-quartz alaskite having allotriomorphic granular texture

mt - magnetite, biotite-rich alaskite and magnetite quartz rock locally replaces a and fa

q - vein quartz

Numbers refer to sample site for geochronologic study by Mose (see Appendix 1).

Figure 11. Blastomylonite-mylonite-gneiss zone with alaskite near sole of Beartown Mountain slice, Cobble Hill, Stop 5 (east to right).

blastomylonite.

Twenty-five feet below the alaskite highly sheared muscovite-biotite-quartz schist of the Dalton(?) contains abundant F_3 folds having axial planes subparallel to the thrust. A slipline determination from this locality is plotted on Figure 6.

Feldspathic schistose marble (Owm) of the Walloomsac can be seen in the field below. The contact of the gneiss, first on the Dalton, then resting on the Walloomsac, can be traced around the east face of the cliffs where the contact is exposed. The intense foliation at the thrust is parallel to the contact. Return to cars by walking east along cliffs following the exposed fault. Continue downhill to Tyringham.

34.4 Tyringham. Turn right on paved road.

35.9 Road sign at crossroads to Monterey. Continue straight on road to Rt. 23 and Otis.

39.2 Bend in road--past the Tyringham-Otis town line. Park for optional stop. The only parking is on opposite (east) side of road. Walk 500 feet east to low cliffs.

Stop 6. Recumbent folds in paragneiss sequence associated with ancillary thrusts within the Beartown Mountain slice.

A series of northwest-trending gentle northeast-dipping thrust faults have formed in the paragneiss sequence east of Monterey in rocks of the Beartown Mountain slice. Detailed mapping of the calc-silicate and amphibolite units has shown that there is no great separation on these faults but that the strain has been largely taken up by D_3 ductile behavior. Excellent exposure of recumbent and reclined F_3 folds can be seen east of the road, where gray biotite gneiss in a thrust sheet overlies intensely deformed hornblende-plagioclase gneiss. Hornblende in the amphibolite has recrystallized within the blastomylonitic fabric. Slipline data for this stop are shown in Figure 5D.

Continue south.

40.2 East Otis and intersection with Rt. 23. Turn left.

43.7 Intersection of Rt. 8 at Otis. Turn left, north on Rt. 8.

44.9 Turn right into Otis Ski Mobile Club, just before green house.

Stop 7. Exposure contact of Basin Pond slice with Beartown Mountain slice (Otis quadrangle).

The detailed stratigraphic units of the Beartown Mountain slice have been mapped from the Great Barrington and Monterey quadrangles eastward into the Otis quadrangle. Continuity of map units, including a very distinctive paragneiss-calc-silicate stratigraphy, reveals

complex isoclinal and recumbent folds. However, no major detachment zones exist. North of this point, in the East Lee quadrangle, a series of higher slices without the distinctive stratigraphy override the Beartown Mountain slice; in sequence these are: Upper Goose Pond, Basin Pond, Upper Reservoir, and October Mountain slices. The contact between the Upper Reservoir and Basin Pond slices is marked by abundant development of alaskite, biotite granodiorite, and magnetite mineralization that is responsible for the marked northwest-trending magnetic anomaly (Popenoe and others, 1964). At Stop 9 we will make a traverse across these three higher slices.

The contact of the Basin Pond slice with the Beartown Mountain slice has been traced southward to this point.

The contact between the diopside-calc-silicate unit of the Beartown Mountain slice and a very tectonized and granite-infiltrated garnet-biotite granitic gneiss (perhaps Tyringham) is exposed above the brook. Above the sillimanite isograd, small granitic stringers and pods are commonly found in or near thrust faults. The granite layers are foliated with the blastomylonitic (F_3) fabric and locally appear to be folded, although the granite most commonly is intruded parallel to the limbs of the F_3 folds in much the same habit as the thin alaskites seen at Halls Hill. This intense folding with granitic stringers produces a different kind of fold-thrust fabric that characterizes the fault zones in higher grade rocks of the Otis quadrangle.

Slipline data from this locality are shown in Figure 5E.

Log resumes at Rt. 8. Turn left toward Otis and follow Rts. 8 and 23 into Otis.

46.2 Turn left on Rt. 23 at center of Otis. Follow Rt. 23 through East Otis.

50.6 At bend in road to left, 0.2 mi. out of East Otis, turn left onto Algerie Rd. by small white house on left.

50.7 At "T" intersection, turn left, still on Algerie Rd.

54.2 Intersection of Lee and Westfield Rds., continue straight.

55.1 Turn right at dirt road, entrance to Williams granite quarry.

Stop 8. Williams granite quarry (fig. 12).

Between Otis and East Otis, we crossed the axis of a major northwest-trending F_4 synform that downfolds the nested Basin Pond, Upper Reservoir, and October Mountain slices. North of East Otis, is a broad expanse of the Basin Pond slice, composed almost entirely of the distinctive garnet-bearing granitic gneiss unit seen at Stop 7. West of our present location, typical Beartown-Mountain-slice rocks reappear from beneath the Basin Pond slice in the White Lily Pond window. The stratigraphic units and the structural sequence match those last seen west of the major F_4 synform. This suggests that rocks of the

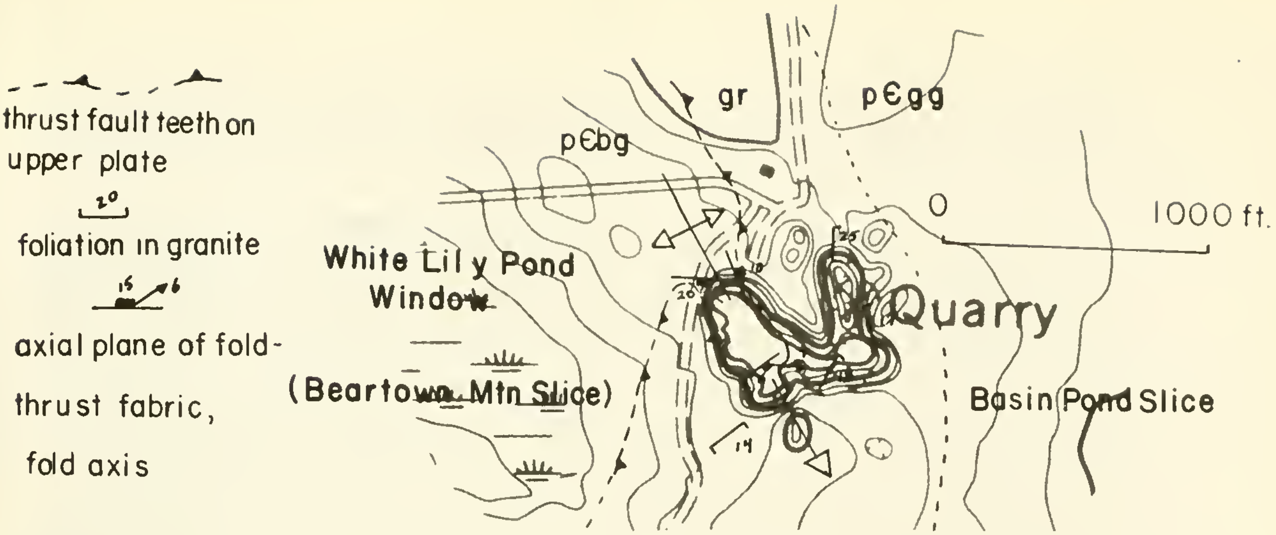


Figure 12. Geologic sketch map of the William granite quarry, Otis quadrangle, Massachusetts. pЄbg=biotite quartz microcline plagioclase paragneiss and amphibolite; gr=gray, biotite-flecked, foliated quartz monzonite of uncertain age; pЄgg=biotite granitic gneiss.

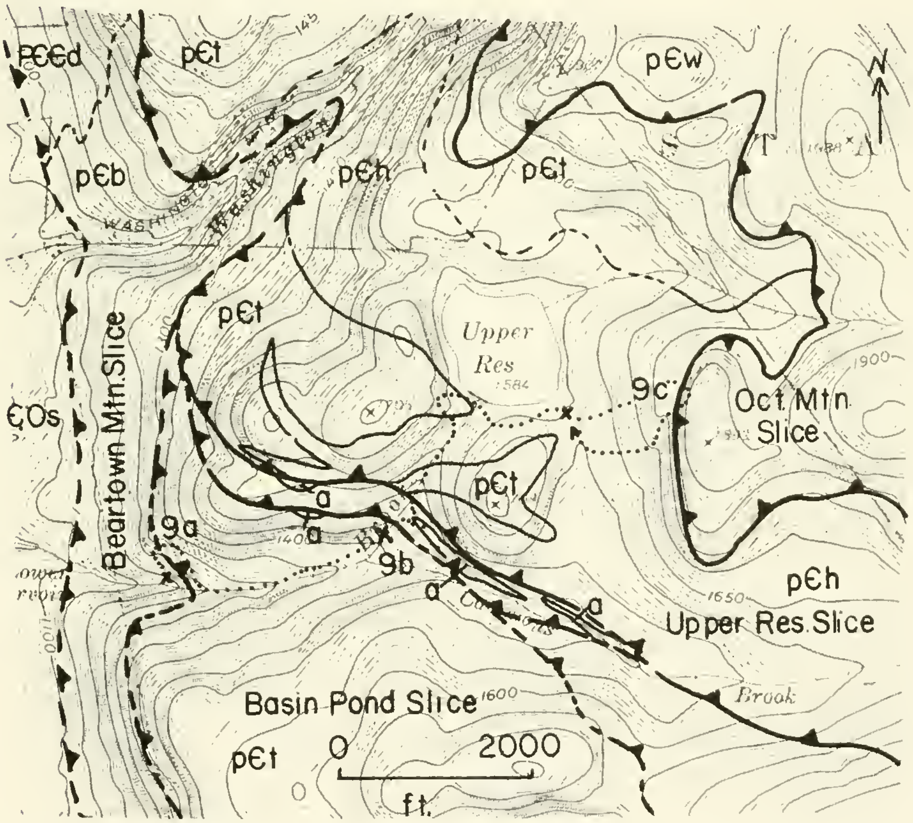


Figure 13. Generalized geologic map of stops 9a, 9b, and 9c, East Lee quadrangle (Basin Pond, Upper Reservoir, and October Mountain slices). pЄw=Washington Gneiss, pЄh=hornblende gneiss, pЄt=Tyringham Gneiss, pЄb=biotite paragneiss, a=alaskite, pЄЄd=Dalton Formation, ЄOs=Stockbridge Formation.

Beartown Mountain slice are structurally continuous beneath the higher slices Throughout this area, dips of the fold-thrust fabric are gentle to the east and west, although northwest (F_4) and northeast (F_5) crossfolds are well developed.

Well-foliated, biotite-spotted quartz monzonite is abundant at or near the contact of the Beartown Mountain slice with the Basin Pond slice, and near the contact of the Basin Pond and next higher October Mountain(?) slice in the northeast corner of the Otis quadrangle. The age, significance, and origin of this quartz monzonite are in dispute at present. I am impressed with the apparent localization of the granite near thrust faults of presumed Ordovician or younger age and believe that the granites may have formed synchronously with the F_3 structures. However, preliminary Rb/Sr studies by Brookins and Norton (1975) gave an isochron age of 605 m.y. $\pm$50 m.y. for samples of similar granite collected from several localities in or near fault slivers in the Middlefield thrust zone. Likewise, preliminary U-Pb analysis of zircon from a nearby quarry also suggests late Precambrian rather than Paleozoic age (R. Zartman, personal communication, 1974).

Rb/Sr whole-rock data by Douglas Mose for granitic rocks collected from four quarries in the Otis quadrangle, shown in Appendix 1, Figure 3, show marked nonlinearity and yield an approximate isochron of 486 m.y. Locally, dikes of biotite quartz monzonite crosscut highly folded country rocks, cross the fold-thrust fabric, and include zenoliths of gneiss. Other exposures of the rock are almost gneissic, having ghostlike traces of compositional layering that is isoclinally folded.

In the quarry, quartz monzonite forms a thin (approximately 90 foot-thick) sheet that dips gently to the east above a highly deformed dark-gray biotite-streaked magnetite-garnet, potassium feldspar-plagioclase-quartz gneiss. Isoclinal, nearly recumbent, reclined folds having granitic stringers intruded subparallel to the limbs of sheared-off folds are common. Zones of blastomylonite crosscut the gneiss parallel to the axial planes, but the quartz monzonite is not similarly sheared. The contact is exposed several places along the west walls and in the bench in the floor of the quarry. Although the gneiss is strongly sheared near the contact, the contact itself is sharp, and the granite appears to be finer grained and possibly chilled at the border.

The conflicting field and isotopic data might be reconciled if the rock we have mapped as a foliated biotite quartz monzonite is anatectic rock remade from 1 b.y.-old gneiss or from Avalonian granite in Paleozoic fault zones.

Log resumes at Algerie Rd.

55.1 Turn right (north) onto Algerie Rd.

56.5 Follow "T" intersection. Turn left.

57.8 Turn left onto Rts. 8 and 20 at Bonny Riggs Corners.

To the west, contacts between the Beartown Mountain, Basin Pond, and October Mountain slices dip moderately to steeply west along the east limb of the large F_4 synform. These west-dipping thrusts have been traced northward by Norton (Trip B-4). Rocks of the Beartown Mountain slice may reappear in the eastern part of the Otis quadrangle in a north-trending elongate window. The structural relationship here suggests that in the Otis and southern part of the Becket quadrangles the rocks of the lowest tectonic slice (Beartown Mountain slice) are exposed in these windows. If the thrust beneath the Beartown Mountain slice is subparallel to the higher slices, the massif may be floored by shallow thrusts even at its eastern margin, and the entire massif may be allochthonous. Rocks of the Hoosac Formation are in thrust contact with the gneisses at the east side of the massif (Norton, 1969, Trip B-4) and would cover the trailing edge of the Beartown Mountain slice, so this problem may never be satisfactorily resolved without deep drilling.

61.0 At bend in Rt. 8 by powerline crossing, we are passing back down through the October Mountain, Upper Reservoir, and Basin Pond slices along the west limb of the F_4 synform.

63.2 Rt. 8 and 20 branch. Follow Rt. 20 to right toward Lee. Hill to right of TV tower exposes stacked slices dipping NE., whereas hills to left contain the Upper Goose Pond and Beartown Mountain slices.

69.2 Branching road just before East Lee Steak House, Maple St. Turn right.

71.5 Follow Maple St. and East St. 2.3 miles to Reservoir St.

71.5 Turn right onto Reservoir St. Log resumes at Reservoir and East St.

Stop 9. (Fig. 13.) This stop consists of a series of substops at the west edge of the massif and will traverse rocks of the Basin Pond, Upper Reservoir, and October Mountain slices (fig. 13). Stop at gate into reservoir and consolidate into most durable high-carriage vehicles, as the road is rough in places.

9a. Park in road near borrow pit just past pump house. View of the Stockbridge Valley to the west, underlain by autochthonous rocks. Dalton of Rattlesnake Hill forms an outlier to west, and in the distance are ridges underlain by allochthonous Taconic rocks of the Everett slice. We are standing on Beartown Mountain slice below the overlying Basin Pond slice. Cliffs to north are highly sheared Tyringham Gneiss of the Basin Pond slice. Gently east-dipping fold-thrust fabric, nearly recumbent folds having a strong blastomylonitic fabric are sheared off along a 10- to 20-cm-thick zone of jet black blastomylonite. Dikes of bull quartz and pink alkali feldspar intrude the shear zones and show a strong lineation that agrees closely with the slipline from separation angle of F_3 folds (fig. 5F). Return to cars.

Continue up dirt road to large cleared area on right.

9b. Optional stop. Exposures of shear zone and alaskite in fault zone at contact of Basin Pond and Upper Reservoir slice. In the field can be seen intensely sheared Tyringham Gneiss having a gently to moderately northeast-dipping blastomylonitic foliation. Fine-grained, mica-poor alaskite locally crosscuts the sheared gneiss and at other localities is foliated with the blastomylonite. This suggests that the alaskite was generated or intruded at the time of the shearing. To the southeast, pods of alaskite locally rich in magnetite are found in this fault zone. This magnetite mineralization is responsible for the sharp aeromagnetic anomaly shown by Popenoe and others (1964). Return to cars and drive across reservoir following dirt road to spillway overflow. At east end of Upper Reservoir.

9c. Rocks at spillway are well-layered biotite, hornblende, and garnet gneiss distinct from the Tyringham. The general trend of the Precambrian gneissosity is approximately east and vertical. This is cut by a gently northeast-dipping blastomylonitic foliation much less intense than at 9b.

Walk northeast to clearing in which hornblende, spotted and streaked garnet biotite gneiss, and amphibolite crop out. Continue northeast over first small ridge which has well-layered hornblende gneiss to a marked north-trending bench, which marks the trace of the thrust zone beneath the next higher slice (the October Mountain slice). Small crops and float of amphibolite and amphibolitic gneiss are found in the bench. Climb the slopes to the east to see the rusty-weathering rocks of the Washington Gneiss. At the base of the slope, blastomylonitic foliation dipping 30° E. has largely obliterated the Precambrian gneissosity. Farther up the slope are blue quartz ribs and distinct beds of the Washington Gneiss.

A slipline of N. 80° E. (fig. 5G), using the rotation sense of minor folds, agrees closely with the N. 75°-80° E. lineation seen on the undersides of blastomylonite zones.

After examining the Washington Gneiss, begin to return to the cars by walking south along the bench following the Washington contact.

About 300 feet south, where the bench terminates, are magnificent exposures of the sheared rocks at the top of the Upper Reservoir slice.

Nearly recumbent folds of biotite hornblende gneiss have a blastomylonitic axial-planar foliation that dips 25°E. Fifty feet east of this crop are exposures of blastomylonite containing small augen of pink feldspar in an even-textured gray foliated blastomylonitic matrix having strong lineation, outlined by pink potassium feldspar streaks; the streaks plunge N. 75° E. at 30°.

71.6 Turn right onto East St., and in 0.1 mile turn left again onto Bradley.

71.8 In 0.2 mi. bear right onto Columbia.

73.0 Intersection Columbia and High St. Turn left.

73.3 Turn left into parking lot of Lee Junior High School.

Stop 10. The Pinnacle - St. Mary's School, Lee. Dalton of the Monument Mountain-Rattlesnake Hill slice.

Tan-weathering muscovite, boitite-flecked feldspathic metasandstone of the Dalton Formation crops out in large cliffs overlying the Cheshire Quartzite seen at the edge of the parking lot. The distinctive spaced foliation seen in the Dalton expresses the fold-thrust fabric. Minute folds of an earlier foliation can be seen as isolated noses of rootless folds. This rock is illustrated in Figure 4 of Ratcliffe and Harwood (1975).

Slipline from separation angle of minor folds and lineations taken from 100-foot section at the base of the cliff is shown in Figure 5I. The slip direction agrees fairly well with the well-developed but variable lineation produced by streaks of pre-existing bedding and foliation in the plane of the thrust fabric.

Muscovite, biotite, granulated quartz feldspar, and recrystallized quartz are oriented in the blastomylonitic foliation. Larger recumbent and reclined folds of quartzite can be seen near the crest of the hill.

The fold-thrust fabric postdates an older schistosity. The Paleozoic rocks in the Berkshire massif were metamorphosed before emplacement. In addition, the faults override and truncate metamorphic (M_1) fold structures in the autochthon. Similar textures and structures are found in the rocks of the Monument Mountain slice.

To return to Great Barrington and Stockbridge, turn left on High St. In 0.3 mile, turn left onto Rt. 20. Follow Rt. 20 1.1 miles to intersection with 102. Turn right. Follow 102 to Stockbridge and take Rt. 7 south out of Stockbridge to Great Barrington.

End of Trip

Appendix 1

Rb/Sr WHOLE-ROCK DATA FROM SELECTED GRANITIC ROCKS IN THE BERKSHIRE MASSIF

Douglas Mose
George Mason University

Rb/Sr whole-rock data for the Tyringham Gneiss is shown in Figure 1. The data reveal that crystallization of the granite occurred about 1170 m.y. ago. Geologic data indicate that the Tyringham Gneiss intrudes Washington Gneiss and other paragneiss units but was involved in the intense pre-Dalton dynamothermal metamorphism (see fig. 7 of text); it thus can be considered a pre- to syntectonic granite.

Rb/Sr whole-rock data for the Cobble Hill mylonite zone (Stop 5) are shown in Figure 2. Although the best fit line yields a somewhat reasonable calculated age, the scatter of the data shows that isotopic equilibrium was not reached throughout the zone of cataclasis. At best, the Rb/Sr whole-rock data show that other means must be used to determine the chronology of faulting. The location of samples is shown in Figure 10 (of text).

Rb/Sr whole-rock data obtained from the Williams Quarry, New Whiting Quarry, and Winn Quarry are shown in Figure 3. The best fit line to all the isotopic data suggests that these granites crystallized in Ordovician time. The data are quite non-linear, so the age estimate is very tentative. The magnitude of scatter from a 486 m.y. isochron is itself unusual, and probably indicates that the granite magmas were not isotopically homogeneous. This inhomogeneity could have been produced if strontium which was absorbed from the surrounding Precambrian gneisses and therefore enriched in radiogenic ^{87}Sr was not homogenized with the strontium within the magma.

Data from Stop 8 are shown in Figure A. Further isotopic investigation will be necessary before the age of these granites is finally resolved.

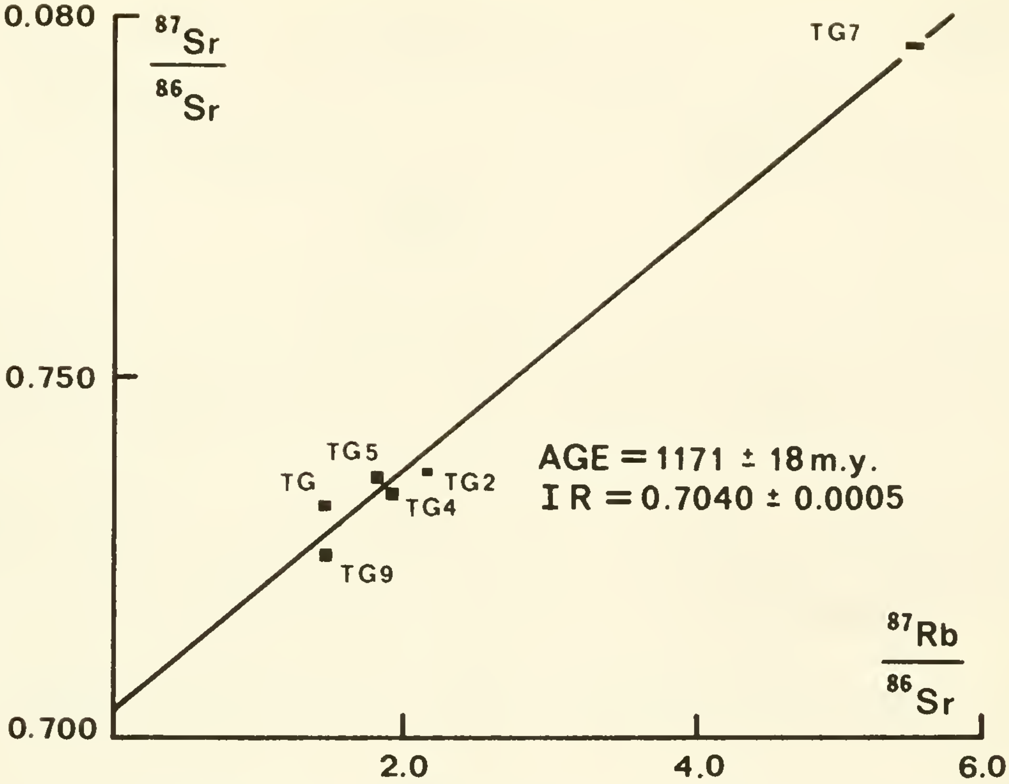


Figure 1. Rb/Sr whole-rock data for Tyringham Gneiss, Monterey, Great Barrington and Stockbridge quadrangles, Massachusetts.

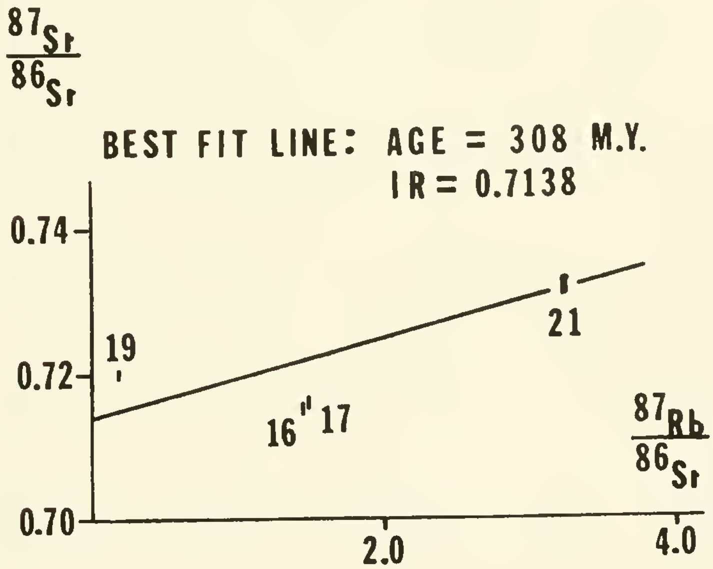


Figure 2. Rb/Sr whole-rock data for alaskite and blastomylonite, Cobble Hill, Stop 5.

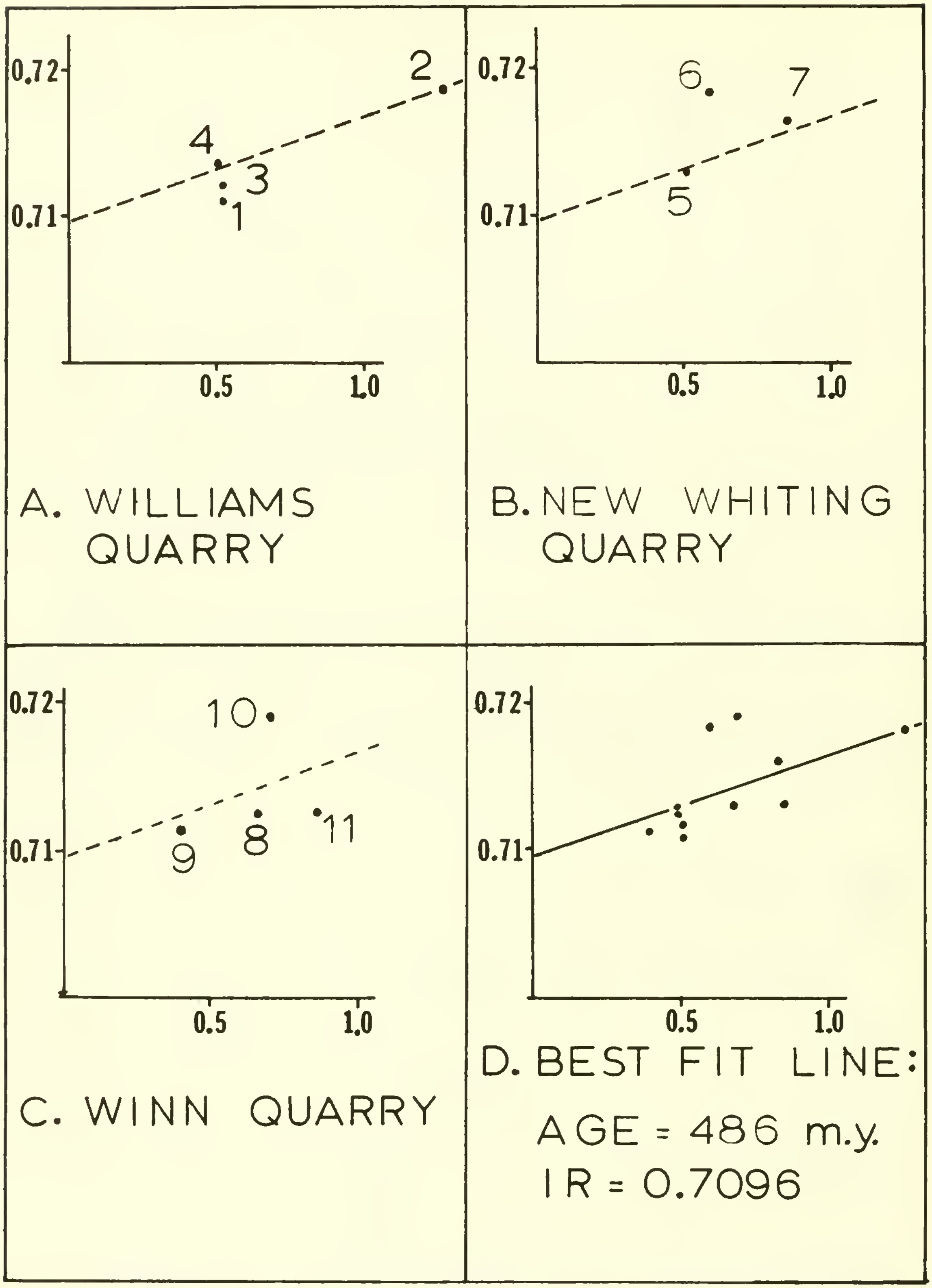


Figure 3. Rb/Sr whole-rock data for biotite quartz monzonite from three quarries in the Otis quadrangle, Massachusetts. A. Williams Quarry, Stop 8. D. shows combined data.

THE LATE QUATERNARY GEOLOGY OF THE HOUSATONIC RIVER BASIN IN SOUTHWESTERN MASSACHUSETTS AND ADJACENT CONNECTICUT

Robert M. Newton,[1] Joseph H. Hartshorn[1] and Walter S. Newman[2]

INTRODUCTION

The latest (Woodfordian) glaciation enveloped the entire region after 25,000-30,000 years ago and stripped the area of most of the weathered mantle left from previous interglacials and the pre-Woodfordian interstade. Directional indicators (whalebacks, drumlins, striations, grooves, boulder fans, and till fabrics) record glacial movement from the northwest quadrant; this direction diverges appreciably from the generally north-striking belted metamorphic terrane. Emerson (1899) noted this divergence and suggested that glacier movement in southwestern Massachusetts radiated southeast from the axis of the Hudson-Champlain lowland. Kelley (1975) found similar ice-flow directions in northwestern Connecticut and concluded that at least in the latter part of Woodfordian time, glacier movement was out of the Hudson Basin southeastward into the Housatonic Basin. The bedrock thalweg of the Housatonic Valley was overdeepened by glacial scour. Borings reveal more than 100 feet of bedrock closure northwest of South Canaan, Connecticut (Melvin, 1970; Holmes and others, 1971).

Deglaciation by downwasting and stagnation appears to have been rapid, and the area was probably ice free by 13,000 B.P. (Before Present). The pattern of deglaciation resulted in major derangement of drainage. Examples include diversion of the main stem of the Housatonic River from its pre-Woodfordian valley to an adjacent tributary valley immediately to the west. The diversion was caused by the deltaic apron and outwash constructed by the proglacial Konkapot River, which joins the Housatonic Valley from the east near Ashley Falls, north of Canaan, Connecticut. In addition, Warren (1971) has trenchantly argued that Great Falls, near Falls Village, Connecticut, the southernmost stop on this field trip, is almost certainly a postglacial landscape feature that resulted from the diversion of the Housatonic River from an unknown earlier course.

The Great Falls bedrock threshold, at an elevation of about 625-630 feet, dammed a lake in the Housatonic Valley that extended north for about 20 miles from Great Falls to the village of Housatonic, north of Great Barrington (Fig. 1). Data from borings (Norvitch and Lamb, 1966; Melvin, 1970) disclose the presence of more than 100 feet of fine-grained lake-bottom sediment at several localities near Sheffield, Massachusetts. The considerable volume of fine-grained sediment indicates that the Great Falls-Sheffield glacial lake endured for at least several hundred years. The demise of glacial Lake Sheffield (formerly called the Falls Village lake; Holmes and Newman, 1971) was due to three factors: 1) minor erosion at the Great Falls threshold; 2) filling of the basin by lake sediments and valley-

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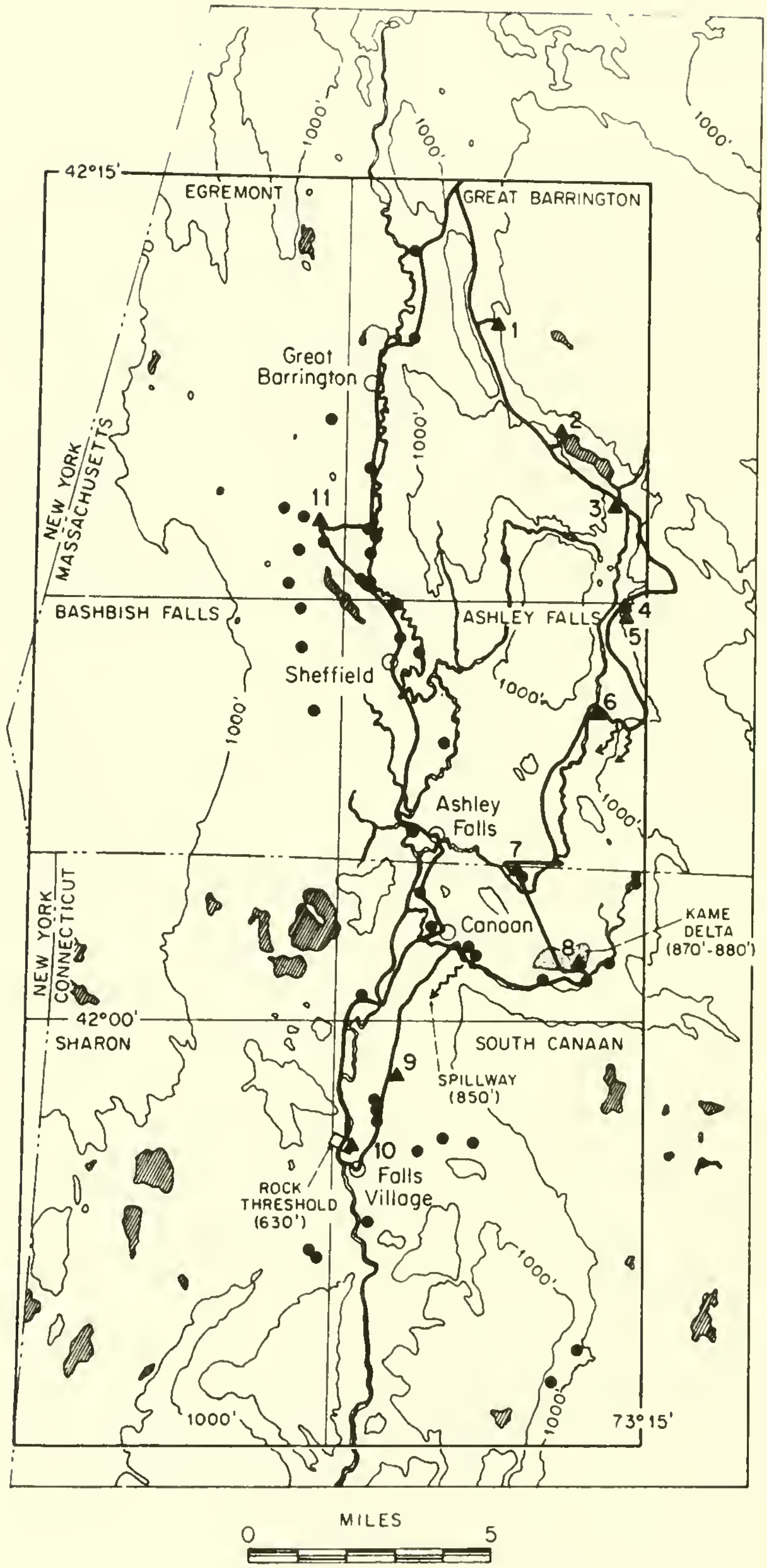
42°15'
EGREMONT
GREAT BARRINGTON
1000'
Great Barrington
NEW YORK
MASSACHUSETTS
1
2
3
11
BASHBISH FALLS
ASHLEY FALLS
4
5
Sheffield
1000'
6
Ashley Falls
7
NEW YORK
CONNECTICUT
Canaan
8
KAME DELTA (870'-880')
42°00'
SHARON
SOUTH CANAAN
9
SPILLWAY (850')
10
Falls Village
ROCK THRESHOLD (630')
73°15'
MILES
0
5

train deposits and early postglacial alluvium grading to the Great Falls threshold; and 3) perhaps differential late-glacial/postglacial isostatic rebound, higher to the north, causing at least partial decanting of the lake basin. The filling of the basin by outwash and alluvium probably buried many proglacial landforms, which may account for the paucity of ice-contact stratified drift in that reach of the Housatonic Valley extending from Great Falls to Great Barrington.

The past 13,000 years witnessed the migration of vegetation into the region, beginning with tundra flora (Whitehead and others, 1973), which was followed by predominantly coniferous forest and then by a mixed conifer-deciduous forest (Whiting, 1974). European Man entered the area during the middle of the 18th century and radically altered the floral composition of the area, while his forest-clearing and agricultural activities increased erosional rates several-fold and perhaps by as much as an order of magnitude.

We find little evidence of pre-Woodfordian Quaternary history in the area. Occasionally, exposures of weathering-profile mantles ("saprolite") are found; the weathered bedrock commonly has a thin cover of relatively unweathered glacial drift. Warren (1971) has attempted to construct an erosional sequence of buried, exhumed, and recently excavated valley segments on a reach of the Housatonic River south of Great Falls. However, without either stratigraphic or chronological controls, it is difficult to accept even his most tentative suggestions.

GLACIAL DEPOSITS

Till

Till forms a fairly continuous mantle over most of the area, except at higher elevations, on steep slopes, and on the summits of some lower hills. Most of the exposed till is sandy, nonindurated, and similar to the upper or younger till elsewhere in southern New England. The clasts are composed of marble, schist, quartzite, quartz, and small quantities of gneiss and granitic rocks. Except for a very low percentage of exotic

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Figure 1. Index map of the Housatonic Valley area in southwestern Massachusetts and adjacent Connecticut. The 1000 foot contour line is shown.

● - Well containing lake-bottom deposits

9 ▲ - Field-trip stop

⊙ - Town or village

- Lake or pond

- Spillway in rock or till

specimens glacially transported from the north and east, the clasts appear to be of local derivation. The till usually ranges from 5 to 10 feet in thickness, although logs of borings document thicknesses greater than 100 feet. These appreciable till thicknesses raise the possibility that till of more than one age may be present.

Stratified Drift

Wastage of the ice in the Housatonic Basin took place in several stages, as recorded by a series of stagnant ice features in tributary valleys. Melting of the ice in the main stem of the Housatonic Valley did not form ice-contact deposits on the valley sides, or, if these features were formed, they were eroded or buried by lake sediments and alluvium. Shortly after the higher elevations became ice free, stratified drift was deposited in glacial lakes and streams adjacent to and in front of the ablating ice. These deposits may be divided into chronological groups on the basis of geographic position and altitude. These chronologic groups of sequencies are successively younger northward and mark retreatal positions of the edge of the ice sheet. In our area, most of the retreatal positions probably were determined by the relation between deposition and topography and very likely do not represent climatically controlled stillstands. The altitudes of deposits generally were controlled by local and temporary base levels, such as glacial ice, bedrock spillways, or glacial lakes. Deposits laid down in association with a stagnant ice margin include such morphological forms as ice-channel fillings, kames, kame terraces, and kame deltas.

Evidence on the relationship of a valley ice tongue in the Housatonic Valley to the uplands and tributary valleys on each side is contradictory.

Perhaps the best evidence that the tributary valleys were deglaciated prior to the disappearance of ice in the Housatonic Valley is found in the Blackberry Valley, east of Canaan. Near East Canaan on the north side of the Blackberry Valley (STOP 8), a huge kame delta was bult into a proglacial lake impounded by an ice tongue in the Housatonic Valley to the west. The topset-foreset interface in the East Canaan kame delta has a reported elevation of 883 feet (Holmes and Newman, 1971), but the delta topography suggests a lower level, perhaps 870-875 feet. The threshold of the lake may have been over the bedrock notch just south of Church Hill, 1 mile south-southeast of Canaan. The present elevation of the Church Hill threshold is about 855 feet.

On the other hand, the Konkapot River valley train seems to show that the ice was gone from the main valley before it left the adjacent tributaries. A well-marked head of outwash (that is, the place where proglacial deposits first become a continuous feature) starts south of Konkapot at an elevation of about 725-730 feet. The valley train declines southward to 690 feet at Clayton, turns westward, and grades into the presumable lake-bottom deposits, or perhaps even post-lake alluvium, at an altitude of about 685 feet. Thus it seems that a body of ice remained near Konkapot after the glacier left the adjacent Housatonic Valley.

RADIOCARBON CHRONOLOGY

We have few data as to duration of the latest glacial interval within the area under consideration. R.L. Melvin (U.S. Geological Survey, Hartford) secured two ^{14}C dates on peat balls from kames near Norfolk, Connecticut, 8 miles east of Canaan. These dates are 28,000±100 years B.P. (W-2043) and greater than 33,000 years B.P. (W-2174) (Sullivan and others, 1970). Newman secured a peat ball from the East Canaan kame delta, which dated at greater than 40,000 years B.P. (W-2615). The East Canaan peat ball yielded no pollen. These three dates suggest that the advancing last Wisconsin glacier incorporated frozen peat during its advance and that these peat balls were deposited in ice-contact deposits. Perhaps the dates indicate an interstade that lasted from about 40,000 to more than 28,000 years B.P. The one finite date in the sequence might indicate that the last glacial ice invaded the area after 28,000 years B.P.

We know of only two early postglacial ^{14}C dates adjacent to our area. Both are basal peat dates from bedrock basins and indicate a minimum age of deglaciation. One of these dates is on material collected by Kelley (1975) in the Ellsworth quadrangle south of our area and was dated at 12,750±230 years B.P. (RL-245). The second date was secured by Whitehead and others (1973) from Berry Pond, west of Pittsfield, Massachusetts, and 20 miles north of Great Barrington. The date at the base of the Spruce Zone was 12,680±480 years B.P. (OWU-481). How long before these dates deglaciation actually took place is moot: we guess about 13,000 B.P.

ACKNOWLEDGEMENTS

Many geologists from the U.S. Geological Survey, both present and former members, have tried in recent years to decipher the complex glacial and deglacial history of southwestern Massachusetts. J.H. Hartshorn and the late G.W. Holmes started reconnaissance mapping in 1962, and Holmes later produced many open-file maps. Detailed mapping in the area covered by this field trip was done by Holmes and W.S. Newman, the latter assisted by R.M. Newton in 1971. We still disagree among ourselves about landforms, sedimentary history, and chronology, and obviously more detailed work of many kinds is necessary before we can decipher a coherent story.

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ROAD LOG

Mileage

0.0 Assemble in parking area of Monument High School at intersection of Route 7 and Monument Valley Road, 3.8 miles north of Great Barrington. Exit left from parking area (enter Great Barrington quadrangle).

0.2 Turn left on Monument Valley Road.

0.9 The road is high on the side of a mined-out kame and kettle complex on right. Konkapot Brook valley to the left.

1.4 Madden Farm, view of ice-contact slope on the left.

2.2 Intersection with Stony Brook Road.

3.1 Turn left on Blue Hill Road.

3.3 Cross Muddy Brook

3.6 STOP 1 - BLUE HILL ROAD KAME TERRACE. Short walk down dirt road on right to exposure in kame terrace. Section exposes colluvium (?)/ eolian/fluvial sediments/lacustrine sediments/gravel. Farther down road is small pit showing the eolian mantle (with ventifacts?)/ channeled fine to medium sand. Note the topography of the kame terrace.

4.1 Return to Muddy Brook Road and turn left (south).

4.8 Kames on left; bedrock and till with boulders on right. Three Mile

Hill directly to the west (right) strikes north-northwest and striations at its summit more or less parallel that same direction, suggesting bedrock control of latest ice-flow directions.

5.8 Intersection with Route 23. STOP! CAREFULLY cross Route 23 (slight jog to the right) and continue south on Lake Buel Road. Note bedrock on right.

6.5 Turn left at Deerwood Camp Road.

7.0 Park along road.

STOP 2 - LAKE BUEL ICE-CHANNEL FILLINGS. This morphological feature is one of several ice-channel fillings found near the shore of Lake Buel. Exposures of the interior of these landforms disclose stratified coarse sand, gravel, cobbles, and boulders. Evidence of collapse is conspicuous.

7.5 Return to Lake Buel Road and turn left. Note till above the break in slope, and many boulders on surface.

8.4 Left on gravel road. Ridge on right is ice-channel filling. Road curves right, crossing through the esker, then swings left and right along the east side of the esker. Lake Buel on left. Bear left along lake.

9.3 Large cut in ice-channel filling on right. Note width and flattish top.

9.7 Rejoin Lake Buel Road and turn left.

10.1 Bear slightly left on Adsit Road. Note channel on right.

10.5 Bear right on dirt road.

10.7 Straight on Crosby Road. Note kettles on right and gravel in roadcuts.

10.8 STOP 3 - CROSBY ROAD ICE-CONTACT DEPOSITS. Park as close to the right shoulder of road as possible. Here we find a large kame immediately to the north; stratified drift containing a large boulder (12 ft x 8 ft x 7 ft) is exposed on the south side of road. A large kettle is present just west of the small pit.

11.0 Cross Konkapot River.

11.3 Intersection. Turn right on Rt. 57. Briefly leave Great Barrington quadrangle and enter Monterey quadrangle.

13.1 Right on New Marlborough Hill Road (dirt road). Reenter Great Barrington quadrangle.

14.6 Join Hartsville Road on right. Leave Great Barrington quadrangle and enter Ashley Falls quadrangle. Here we drive along the base of a

high and complex kame terrace.

14.7 STOP 4 - HARTSVILLE ROAD TILL. Till at base of kame terrace is loose, sandy, and light colored, and has some medium to fine sand laminations within the matrix. Clasts are dominantly quartzite and include some well-weathered gneiss. Note flood plain of Konkapot River on right side of road.

15.0 Left on gravel road up past houses and barns.

STOP 5 - HARTSVILLE ROAD KAME TERRACE. Large pit at base of kame terrace. At southern end of pit slumping is prominent along the ice-contact face of the terrace. Material is predominantly medium to fine sand. Climbing ripple cross-laminations indicating south-flowing currents occur in the upper part of the exposure. Also present are several beds of pebble to cobble gravel in which low-angle crossbedding dips south. Pebbles are mostly quartzite, marble, and gneiss.

15.3 Return to Hartsville Road.

15.6 Intersection with Great Barrington Road.

15.8 Intersection with Hayes Hill Road.

16.0 Left on Mill River Road.

17.1 Intersection with Lumbert Cross Road. Bear left onto Lumbert Cross Road.

17.8 Intersection. Cross onto Rhodes and Bailey Road.

17.9 Thick eolian mantle in roadcut on right.

18.2 Bear right onto Cagney Road.

18.6 Note gorge, an ice-marginal channel, on left just before sharp right turn in road. Entrance to gorge is 1030+ feet.

18.8 A second lower gorge is down the steep slope to the left between road and next hill. Small pond may be seen within the gorge.

18.9 Intersection with Southfield Road. The intersection is at the upstream end of the lower Cagney Road gorge, whose elevation is 890+ feet.

19.0 Turn sharp right over bridge, then sharp left on Hadsell St.

19.5 Left turn into New Marlborough Township Park. Park at far end of parking area. Bring lunch!

STOP 6 - UMPACHENE FALLS. The stratigraphic section here at the falls is reversed; the younger Cheshire Quartzite crops out beneath

the older Cambrian Dalton Formation. As you ascend the falls, you go down section stratigraphically.

UMPACHENE FALLS SAPROLITE. Proceed to north end of parking area adjacent to Konkapot River. Outcrop of deeply weathered Stockbridge Formation. Approximately 10 percent of all observed bedrock exposures in the field-trip area show deep weathering profiles.

19.6 Left on Hadsell Road. Cross Konkapot River on rickety single-lane bridge.

19.7 Left on Clayton Mill River Road.

20.5 Bedrock outcrop on left. Distorted marble. Note resemblance to kame on the topographic map.

21.0 Rise onto bedrock-cored hill. Note knobs of bedrock off in fields to left. Descend towards Konkapot.

21.2 Right turn at Konkapot and continue on Clayton Mill River Road. Note head of outwash to left at 720+ feet. Continue south on Clayton Mill River Road on small kame terrace that is part of the head of outwash.

21.9 Kettle on left in head of outwash. The meltwater streams coursing down the valley, in part over and around isolated blocks of ice in its northern part (note kettles and swell and swale topography), built a valley train between the hills on either side and surrounded a kame (present elevation 759 ft; possible original height 80 ft or more). The surface of the outwash declines to 690 feet at Clayton, and the gradient flattens as the outwash turns the corner toward Ashley Falls.

22.3 Intersection with Alum Hill Road. Note coarse gravel in fields on right.

23.0 High kame belonging to earlier sequence on left.

23.4 Intersection with Canaan-Southfield Road. Continue south.

23.6 Village of Clayton. Turn right on Old Turnpike North. As we go west, note the small terrace at 700+ feet, which is higher than the Konkapot valley train. The road goes up on to the terrace for a distance and descends again to the valley-train level.

24.7 Turn left on Allyndale Road.

25.2 Left turn into gravel pit.

STOP 7 - SODOM GRAVEL PITS. These pits in pebble and cobble gravel are approximately 2 miles south of the head of outwash north of Clayton. This alluvial and/or deltaic complex deposited sediment completely across the Housatonic Valley and is still responsible for

the impingement of the Housatonic River against the marble on the west side of the valley. Indeed, the outwash apron probably completely buried the bedrock highs in and about the village of Ashley Falls and the sediment completely surrounded several other bedrock knobs in the vicinity. After deglaciation, the Konkapot cut through the outwash sediment and into the underlying bedrock, creating Ashley Falls. Note eolian filling in shallow swale in gravels. Ventifacts (?) are rare. Exit from gravel pit and turn left on Allyndale Road.

26.0 Sodom! Continue south on Allyndale Road. Cross Konkapot River and Squabble Brook; note fine-grained alluvium. Note alluvium/till contact at base of hill.

27.2 Enter onto surface of East Canaan kame delta.

27.4 Left turn onto dirt road just beyond entrance into O'Connor Bros. gravel pit.

STOP 8 - EAST CANAAN KAME DELTA. Topset beds of gravel and foreset beds of fine to medium sand are exposed in the many pits. Kettles and ice-contact stratified drift occur at the north end of these pits. Peat balls are occasionally found in the foreset beds. The delta was built into an ice-dammed lake having its level at about 870-880 feet. The lake was presumably held in by ice blocking the valley to the west.

28.6 Turn 180° to right onto Canaan Valley Road. Face of delta on right; flood plain of Whiting River on left.

29.5 Right turn on Route 44 West.

30.1 Village of East Canaan. Turn left onto Lower Road.

30.5 Intersection.

31.9 Limestone quarry on left.

32.2 View left of possible spillway of the glacial lake associated with the East Canaan kame delta. Elevation of Church Hill spillway is 850 feet (too low for lake outlet?).

32.8 Turn left (south) on Route 7.

33.3 Right on Sand Road.

33.5 Railroad crossing.

34.6 Bear left at intersection.

35.6 STOP 9 - SAND ROAD SAND DUNES. This exposure of fine sand with little internal structure suggests eolian deposition. Springs and drainage nearby seem to emerge from beneath a crust on saprolite.

36.7 Join Route 126. Continue straight on across railroad tracks.

38.3 Sharp 120° right turn under railroad overpass past electrical powerhouse.

38.6 Bridge over Housatonic River.

38.7 Bear slightly right.

38.8 Right turn onto Housatonic River Road.

39.3 STOP 10 - GREAT FALLS. Before construction of the dam in the early 20th century, the cataract possessed a clear fall of approximately 60 feet. Warren (1971) stated that the Housatonic was diverted across this bedrock sill sometime during the Wisconsinan Stage. This sill exerted a powerful influence on the late-glacial and postglacial landscapes. First, the Great Falls sill ponded a lake that ultimately extended north about 20 miles to the vicinity of Great Barrington, Massacusetts. Second, the Housatonic south of the falls is generally confined to a narrow gorge, but towards the north it is a meandering stream in a rather wide valley. Here we have a classic example of a local base level controlling the regime and grade of a segment of a major river. Look for "meteorite."

39.4 Road passes through short ice-marginal channel.

40.9 Alluvial-fan complex.

41.8 Leave South Canaan quadrangle; enter Ashley Falls quadrangle.

42.1 Turn right on Route 44.

42.7 Note oxbow lake on right.

42.9 Cross Housatonic River and bear left on Route 44.

45.2 Village of Canaan. Turn left toward North Canaan Town Hall.

46.9 Canaan Town dump.

48.0 Intersection with Route 7A. Turn left.

48.2 Village of Ashley Falls. Turn left at red light on Andrus Road.

49.3 Cross Housatonic River. Note outcrops exposed where Housatonic impinges against bedrock knob known as Bartholomew's Cobble. The Cobble is a small bedrock "island" surrounded by alluvium. Also note meanders on right.

49.4 Turn right on Andrus Road.

51.0 Dangerous intersection, cross railroad tracks. CAUTION. Turn left on Route 7A.

51.5 STOP. Left turn on Route 7.

53.5 Sheffield. The sediments of glacial Lake Sheffield achieve their greatest thickness in the vicinity of this village. These lacustrine sediments are overlain by valley-train sands and finer alluvial sand. Continue north on Route 7.

54.9 Leave Ashley Falls quadrangle and enter Great Barrington quadrangle. Prepare to make left turn.

55.1 Left turn onto Egremont Road.

55.2 Cross railroad tracks.

56.0 Road leaves alluvium.

56.2 Leave Great Barrington quadrangle and enter Egremont quadrangle.

56.9 Intersection with Limekiln Road.

57.2 Right turn onto gravel road. Park on right shoulder of road.

STOP 11 - SHAY'S REBELLION (1786-1787) THE LAST BATTLE. The site of the end of Shay's Rebellion. Visit Monument. Small kames suggest ablating ice whose meltwater may have passed through the gap to the east. Continue east through gap.

57.6 Limekiln Gap.

57.8 Leave Egremont quadrangle and reenter Great Barrington quadrangle. Cross high-level terrace and descend to modern alluvium.

58.3 Intersection with Route 7. STOP! Left turn on Route 7.

61.3 Intersection with Route 23. Continue north through Great Barrington.

62.2 Right turn on Route 7 on bridge over Housatonic River.

62.7 Bear left (north) on Route 7.

66.4 Right turn into Monument Mountain High School parking area.

END OF TRIP

THE GLACIAL GEOLOGY OF THE HOUSATONIC RIVER REGION IN NORTHWESTERN CONNECTICUT

George C. Kelley

A recent (Kelley, 1975) investigation of Late Pleistocene and Recent surficial deposits in western Connecticut determined characteristics of Wisconsin glaciation and the history and chronology of deglaciation in part of the dissected New England Uplands.

This region lies along the midreach of the Housatonic River in western Connecticut, and has local relief exceeding 1,200 feet. Surface morphology and internal characteristics of glacial and glaciofluvial erosional and depositional features were examined and mapped in detail in the Kent and Ellsworth, Connecticut, U.S.G.S. 7-1/2 minute quadrangles, and by reconnaissance in the adjacent quadrangles.

Ice along the lateral east margin of the southward-waxing, Wisconsin-age Hudson-Champlain Valley ice lobe successively overran ridges trending northeast-to-southwest. Late Wisconsin ice flow was consistently toward the southeast in this area. Glacial erosion on the upland surfaces was weak, and several early or pre-Wisconsin melt-water channels persist, which evidence little late Wisconsin glacial or glaciofluvial modification. Deeply weathered rock has been preserved beneath unweathered till. Till deposits are generally thin, averaging from 10 to 15 feet in thickness, but some till deposits exceed 200 feet thickness. Direct evidence for two or more cycles of till deposition is lacking, although multiple glaciations can be inferred from drainage derangement of the Housatonic River and from anomalies in configuration of old, upland melt-water channels which were re-occupied and eroded by melt water during subsequent deglaciations.

The orientation of ridges and the local terrain relief exerted minor control on ice flow during waxing phases of glaciation. Local relief and ridges which were oriented transverse to ice flow became the dominant control factors for ice flow during late phases of deglaciation and ultimately initiated marginal stagnation zones.

Late Wisconsin deglaciation evolved in three stages in this region. First, the active ice margin receded rapidly northwestward across, and almost transverse to, the upland ridge crests in response to factors of both backwasting and downwasting. Second, local terrain relief restricted active ice flow, initiated stagnation, diverted melt-water flow and controlled deposition of small active ice-marginal deposits on the northwest slopes of ridges. Third, melting and thinning of stagnant ice tongues in valleys with ice surfaces which were low gradient and southward-sloping caused rapid northward recession of the stagnant ice margin. Sequences of related outwash deposits have been correlated with inferred ice-marginal, recessional positions. The zone of stagnant ice distal to active ice ranged from 6 to 15 miles in average width.

Lacustrine sediments accumulated as stagnant ice block melted in isolated basins and other depressions where through-flowing melt-water drainage was restricted or absent. The paucity of ice-contact and outwash deposits in the isolated basins indicates that little entrained debris was present in the stagnant ice. Prograding outwash along the Housatonic River and other major drainage routes infilled glacially overdeepened rock basins and buried underlying lacustrine sediments beneath upward-coarsening sand and gravel.

Multiple glaciations in the Housatonic River region of northwestern Connecticut are evidenced by till diversion of the Housatonic River drainage near West Cornwall; anomalous configurations of melt-water channels on the

uplands at East Kent and near Mohawk State Forest; the disparity of erosional incisions in the Stockbridge marble at Great Falls and Bulls Bridge; and till retention in the New Preston melt-water channel.

Erosional competence of the last glacial ice may have been weak in this region, as indicated by several wells which contain weathered rock beneath till; several small, pre-Late Wisconsin, upland melt-water channels which evidence only minor glacial modification; and the lack of excavation of stream-diverting till near West Cornwall.

The latest active ice flow across this region was part of the east lateral margin of the Hudson-Champlain Valley ice lobe, as supported by the consistent regional trend from northwest to southeast (S. 40° E.) of ice-flow indicators including striations, drumlin axes, and stoss-and-lee topography; and the regional trend of upland melt-water flow towards the southeast from thicker ice to the northwest.

High topographic relief controlled waning glacial ice flow which caused leeward effects responsible for Late Wisconsin ice-flow diversions north and east of this region. This is supported by the absence of north-to-south and northeast-to-southwest striations and secondary drumlin tails which are present in nearby regions.

Deglaciation in this region was initiated by recession of an active ice margin from southeast to northwest, as evidenced by active ice-marginal till thickening and associated outflow melt-water drainage on the ridge southeast of the East Aspetuck River valley; and the persistence of thick ice and associated deposits northwest of ridges versus associated but stagnant ice-related features southwest of each ridge.

Striations and the orientation of erosional remnants of melt-water drainage indicate the receding active ice margin was transverse to the trend of the major ridges.

Scattered, small, ice-marginal deposits and fragmentary drainage channels preclude correlation of definable, upland, active ice-marginal stillstands, and thus support the concept of a generally rapid, active ice-marginal recession across the uplands.

Ice thinning resulting from backwasting and downwasting eventually caused topographic relief to become the dominant factor controlling the local mode of deglaciation. This is evidenced by the successive transition from weakly active ice-marginal features to stagnation morphologies where ice flow to the southwest was restricted by the upland ridges. Active ice continued to flow through broad, low-altitude cols during late phases of deglaciation after flow across the uplands terminated, as evidenced by the development of valley plugs south of Gaylordsville near the Dogtail Corners col, and north of the Tarradiddle near the Lakeville col; and drainage diversions, linear boulder concentrations, and associated melt-water drainage features east of Gaylordsville.

Low-gradient, ice-contact outwash deposits indicate that downwasting of stagnent ice tongues and blocks dominated deglaciation in the narrow valleys and semi-isolated basins following the restriction of active ice flow by adjacent uplands.

Small, fragmentary outwash heads and small, vertical intervals between kame terraces record the gradual change of terminal and marginal position of stagnant ice tongues. These changes were in response to the interaction of the ice-surface gradient and the configuration of the underlying terrain as stagnant ice downwasted.

The limited areal extent and thickness of ice-contact sediments associated with disintegrating ice blocks in the semi-isolated basins and valleys which

were restricted from through melt-water flow, and the general regional thinness of till, suggest that entrained rock debris was sparse in the ice masses overlying this region.

Incised melt-water channels and chutes, and the position, textural composition, and internal and external characteristics of glaciofluvial deposits indicate that melt water moved on, along, in and beneath active and stagnant ice as melt-water flow toward the southwest frequently changed its drainage routes.

Lacustrine silts and varved clays define small, ephemeral ponds which developed in the narrow Housatonic River valley and in semi-isolated basins during late phases of deglaciation. Larger valleys to the west and areas with northward drainage developed larger, persistent ponds into which prograding outwash extended and buried or partly buried underlying lacustrine materials.

Upland bogs, which have developed postglacially, contain as much as 22 feet of organic material mixed with silt and clay. An isotopic radiocarbon date (RL-245) and preliminary pollen analysis made on samples of buried peat indicate that the Housatonic Highlands were free of ice and that forest vegetation was becoming re-established by 12,750 $\pm$ 230 years B.P.

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ROAD LOG

TOPOGRAPHIC QUADRANGLE (1:24,000, 7-1/2 minute) - Great Barrington, Ashley Falls, South Canaan, Sharon, Cornwall, Ellsworth, Kent, Dover Plains

0.0 From Monument Mountain School proceed south on Route 7 through the towns of Great Barrington, Sheffield, Ashley Falls, Canaan, and South Canaan, approximately 22.7 miles. Observe the flat, low-gradient valley floor on which the Housatonic River meanders. Sand, which is exposed in many places, overlies thick lacustrine clay and silt.

22.7 Turn right (northwest) onto the road entering Falls Village. Continue through business district to intersection at railroad tracks.

23.2 Turn right (northwest); continue about 200 feet; turn left; cross tracks.

23.6 Bear left; cross bridge; keep right, following the Housatonic River.

24.0 STOP 1 - GREAT FALLS
The Housatonic River flows over Stockbridge marble at Great Falls. Holocene nickpoint migration of about 20 feet, and stream incision of 10 to 15 feet, are evident. The falls served as a rock threshold which controlled ponding (Glacial Lake Great Falls) and sediment deposition in the valley north of the falls during deglaciation (Holmes and others, 1971). It is upon these sediments that the modern Housatonic River meanders. There is presumably a preglacial, graded stream channel from which the ancestral Housatonic was diverted by deposition of thick till. About two miles south of Great Falls, the Housatonic River enters the narrow Housatonic gorge, which is incised in gneiss of the Housatonic Highland.

Turn around, proceed south, re-cross the Housatonic River. Bear right, and continue toward the railroad tracks, but do not cross them.

24.8 Turn right (Appalachian Trail). Proceed southward past the regional high school. The terrace surface to the right, across the Housatonic River, has an altitude of approximately 610-620 feet.

26.3 Intersection. Cross Route 7 CAREFULLY, and continue southeast.

27.1 STOP 2 - GRAVEL PIT COMPLEX - KAME TERRACES OR KAME DELTA?
Exposures of sand and gravel in these pits are badly slumped, but the large boulders and contorted strata indicate an ice-contact origin. Terrace surfaces at 570, 610, and 650-foot altitudes record different episodes of melt-water deposition. The narrow Housatonic River gorge is choked for 2.5 miles south of these pits with poorly exposed, ice-contact deposits. This stop is at approximately the position of an ice margin (perhaps active), distal to which lay thin, stagnant ice. As the margin gradually retreated through this zone, glacial debris was delivered to the stagnant ice zone beyond. Buried ice, armoured with debris, presumably melted slowly, developing the distinctive knob-and-kettle valley plug. A rapid northward retreat or dissipation of the ice from this plug initiated ponding (Glacial Lake Lime Rock) in the intervening fosse (Holmes and others, 1971). Salmon Creek (Sharon quadrangle), a tributary to the Housatonic River, delivered debris-laden melt water to this pond, and built a delta with a surface altitude of 590-600 feet. The delta today is the site of the Limerock Racetrack. Terrace deposition, to altitudes approximating 600 feet, were controlled by the damming effect of the valley plug. Higher planar surfaces record an ice-contact kame terrace or kame-delta deposition. Lower terraces represent postglacial stream terrace development. Excavation of the valley plug materials by the Housatonic may have been rapid, and presumably was partly controlled by the location and melting of buried ice. This reach of the river has

deposits which record the transition from a zone of accumulation in a proximal part of a fluvial ice-contact sequence to a fluvial-lacustrine, non-ice-contact sequence (Koteff, 1974). The valley plug served as a temporary base-level control.

Turn around, proceed north, returning to Route 7.

27.8 Turn left on Route 7; Cross the Housatonic River

28.0 Bear right onto Route 112 at the triangle.

28.9 STOP 2a - LIMEROCK RACETRACK - A VIEW OF DELTA MORPHOLOGY (if racing is not in session) Which terminated first - ponding behind the valley plug, or delivery of debris-laden melt water by Salmon Creek?

Return to Route 112. Proceed east to intersection with Route 7.

29.9 Valley plug knob-and-kettle topography evident for next two miles. Materials include tills and collapsed glaciofluvial materials. Comparable deposits are located 4.5 miles to the northeast in Hollenbeck River valley.

34.0 Turn left (east) onto Route 128. Cross covered bridge and turn right (south) onto secondary road. Proceed to Trinity Church property.

34.5 STOP 3 - WEST CORNWALL - HOUSATONIC RIVER DIVERSION

Evidence indicating diversion of the ancestral Housatonic River includes thick till (225 feet) in the small valley east of the present river course (Warren, 1971). The buried floor of this valley is at least 50 feet below the modern Housatonic, and may have been the first drainage channel. Prior to achieving the modern channel, the Housatonic may have been diverted a second time. Evidence includes an overdeepened segment of the river lying upstream from a short segment flowing on rock. Pine Swamp Brook is barbed at its junction with the Housatonic River, and has an abnormally steep gradient in its lower reach. Warren (1971) advocates two diversions, and speculates that the first was caused when the small valley was plugged by Lower Pleistocene till of Kansan or Nebraskan age. The second diversion, of perhaps Illinoian age, established the modern course. This allowed sufficient time for the Housatonic River to incise almost 140 feet into the resistant gneiss of the Housatonic Highland. Any additional Holocene modification was presumably limited in extent, because the river has incised only 10-15 feet into the less resistant marbles at Great Falls.

35.0 Turn left (south) onto Route 7.

36.3 The Housatonic Meadow State Park campground has several terrace levels and an isolated kame. Some of these are visible from the road - terraces to the right, kame to the left. These record in-echelon, kame-terrace deposition by gradually lowered melt-water streams (Kelley, 1975). Comparable terrace fragments have been mapped on the east side of the river (Warren, in press).

37.8 Turn left onto secondary road north of the intersection of Routes 4 & 7. (USE CAUTION - opposing traffic has right-of-way) Church to the left is situated in one of several naturally formed depressions.

39.9 STOP 4 - KAME TERRACES, KETTLE?, AND STREAM TERRACES

A test well (Melvin, 1970) penetrated 102 feet of sand and gravel. Bedrock contours inferred from well data, rock outcrops, and slope morphology suggest that this reach of the Housatonic is a closed rock basin filled with glaciofluvial deposits. Underlying rock should be carbonate (Rogers and others, 1959). Problem - is the depression a kettle or a sink hole? A small Holocene stream terrace is evident near the depression. A kame terrace flanks the hill slope to the west. Silts mantle some higher altitude ice-contact and till deposits along this reach of the Housatonic River, but lower altitude ice-contact deposits lack this mantle. It is inferred that initial kame-terrace deposition beside stagnant ice tongues

was followed by ponding of waters on and around the tongues. Glaciofluvial deposition was subsequently renewed on and against thinner stagnant ice. Problem - What caused the ponding?

Return to Route 7, turn left.

41.0 Bear left (southeast) onto Routes 4 and 7; cross Cornwall Bridge.

41.2 Flat surface to left of highway is artificial, and originally contained an ice-channel filling with large, imbricated boulders, eight feet in length, which indicated melt-water flow direction from Furnace Brook.

41.3 Bear right (south) onto Route 7.

43.1 Large, rounded erratic at left of road. Silt mantles, kame terraces, and isolated kames to the right of the road.

45.2 Kent Falls State Park.

46.2 Bear right onto secondary road. Till is exposed in road banks.

46.3 Ice-contact gravels are in road banks. Kames and channel-filling features are evident north and south of the road.

46.8 STOP 5 - STANLEY WORKS INC. GRAVEL PIT

A test well (Melvin, 1970) penetrated 74 feet of silts, sands, and gravels beneath this pit. Well data and bedrock exposures in the river channel and on hill slopes indicate that this is a rock basin. Exposures in this pit have changed during excavation. Steeply dipping foreset bedding records the prograding of glaciofluvial materials into ponded water. Valley-train deposits in the central part of this basin are part of a fluvial-lacustrine, ice-contact sequence (Koteff, 1974). Lateral, higher altitude deposits, and large kettles to the northeast, record earlier ice-contact deposition. The texture of materials decreases downstream from boulder gravels to pebble gravels. Abandoned stream channels are believed to record Holocene stream adjustments.

Return to Route 7.

47.4 Turn left (northeast) onto Route 7.

47.9 Turn right (southeast) onto secondary road.

50.0 Turn left (east); proceed about 400 feet to gravel pit entrance.

STOP 6 - WILSON ROAD GRAVEL PIT - OUTWASH HEAD

This upland gravel pit is in an outwash head. Melt waters entered from the north and west. Foreset beds in the upper level materials dip to the southeast. Material texture changes from the underlying coarse, poorly sorted, cobble-boulder gravel to the overlying well-sorted sands. Collapse structures are sometimes evident. The upland area where these materials were deposited is the lowest col crossing the ridge for several miles. Melt-water deposition was into ponded water around stagnant ice blocks. In addition to materials in this pit, kames and kame terraces are present in the small valley west of Wilson Road. A local base level threshold for the pond into which the materials were deposited for this pit has an altitude of about 1,180 feet. The lowest base level in the col is 1,155 feet altitude, 0.4 mile northeast of East Kent. This threshold controlled deposition west of Wilson Road. Impervious ice at this outwash head maintained melt-water flow across the 1,155-foot threshold, while ice in the valley southeast of the ridge thinned. Erosional terraces on till, and related boulder gravels in that valley, record a 780-foot ice-surface altitude at the time melt-water drainage terminated (Kelley,1975). Deposits in this col are part of a lacustrine ice-contact sequence (Koteff, 1974).

Return to Wilson Road and proceed south.

50.6 The 1,180-foot threshold is to the right of the road.

50.9 Turn right (west) onto Route 341.

52.0 Turn left (southwest) onto secondary road.

52.2 STOP 7 - BEAMAN POND - DIRECTIONAL INDICATORS
Striations and chatter marks are well preserved on a glacially polished quartzite surface. Additional striations on surfaces along Route 341 have the same general orientation of S. 25-30° E. Problem - Did stagnant, isolated ice blocks persist in upland basins like those now occupied by the Spectacle Ponds while stranded ice in the valleys rapidly thinned? If so, how did melt-water escape from the Spectacle basins without destroying these striations? Small, deeply incised channels north and east of Beaman Pond appear to have a fluvial origin. Glacial striations, however, have been observed in some of these channels. Other channels start abruptly, but have hanging tributaries extending farther upslope than the main trunk. Warren (personal communication) also observed anomalies in upland melt-water channels to the northeast on this ridge (Cornwall quadrangle). When and how were some of the upland channels formed? It is postulated more than one cycle of upland glaciation is responsible.

52.3 Return to Route 341; turn left (southwest). Thin till with numerous rock exposures is evident on the upland.

55.8 Turn right (northwest) onto Cobble Road. This intersection is in a melt-water spillway (threshold altitude 575 feet) that controlled ice-contact deposition for kames, ice-channel fillings, and a kame plain evident in this small valley. These features belong to a fluvial ice-contact, or to a fluvial-lacustrine, ice-contact sequence (Koteff, 1974).

57.4 STOP 8 - FLANDERS FIELD - BOULDERS
This linear boulder concentration presents a problem. Is it a boulder train, a boulder moraine, a "lag deposit," or simply a typical New England concentration of erratics? Boulders are gneiss, carbonates, and quartzite (all local lithologies). Marble underlies this terrain. Many boulders to the east were buried by an industrious farmer. A kame-delta associated with the ice-channel deposit near the hill to the east closely approximates the original linear extension of these boulders. Melt-water drainage moved east and south of Flanders toward the 575-foot altitude threshold, while thick ice in the Housatonic River channel blocked melt-water drainage toward the southwest. Abandoned channels near Kent Furnace, and the areal relationship of deposits in the Housatonic Valley and the valley through South Kent (Kent quadrangle), suggest that ice thinned more rapidly south of Flanders than at Flanders. At Flanders, ice-surface altitudes of about 580 feet were maintained, while in the valleys 1.25 miles farther south, surface altitudes on ice were lowered to 450 feet.

Proceed 0.2 mile northeast to intersection with Route 7.

57.6 Turn left (west) onto Route 7. Observe boulders in fields and walls.

59.2 Kent is on the surface of a valley train at a 400-foot altitude.

59.5 Weakly incised, abandoned melt-water channels on the surface of the outwash to the left of the highway may be the last remnants of active melt-water streams delivering debris to the valley train. Lower altitude stream terraces are evident farther up the hill slope.

64.1 STOP 9 - SPOONER HILL - ABANDONED MELT-WATER CHANNEL
High-altitude boulder concentrations on the northwest-facing hill slope, and knob-and-kettle topography south of this channel, record the movement of melt water along an ice margin through the channel (threshold altitude 603 feet) to lower altitude ice beyond. Beyond the channel, ice at a surface altitude of 500 feet held the melt water against the hill. Knob-and-kettle topography farther south, with surface altitudes to 450 feet, record deposition by this melt water. Ice-surface altitudes across

Spooner Hill differ by approximately 100 feet through a distance of one-half mile. Was ice west of Spooner Hill "live ice" that moved through the wide col to the southwest in the Housatonic Highland? Is topographic control significant during late phases of glaciation? Thinning ice west of Spooner Hill eventually opened lower altitude melt-water drainage routes to the southwest.

Proceed east.

65.3 Turn right (southeast). This road is in the South Kent valley.

65.6 Ice-contact topography to the right of the road. Swamp to the left indicates that melt-water drainage and deposition around a stagnant ice block terminated before the block melted.

66.8 Knob-and-kettle topography.

67.3 Flat-topped kames to left of road. The largest kame has two distinct, flat surfaces at different altitudes. Why?

67.7 Turn right (northwest) at intersection.

67.8 Make sharp left turn onto Route 7 and cross the Housatonic River.

67.9 Bear right (southwest) onto Route 55.

68.3 Bear right onto small road and enter driveway to gravel pit.

STOP 10 - GAYLORDSVILLE GRAVEL PIT

Deposits contained in the Gaylordsville basin and the Housatonic Valley beyond Strait's Rock indicate that an ice-cored valley plug south of Strait's Rock melted and gradually lowered the threshold to which deposits on and along stagnant ice in the Gaylordsville basin were graded. Minor ponding was followed by stream deposition which partly buried silts and sands beneath poorly sorted gravel. These glaciofluvial gravels can be traced northward along the Housatonic River to its confluence with the Tenmile River, and thence westward (Dover Plains quadrangle). They cannot be continuously traced farther north in the Housatonic River valley. This pit records the ponding, valley-train deposition, and subsequent stream terracing which evolved in this basin.

Return to Route 55

68.9 Turn right (west) onto Route 55. This road is in an abandoned melt-water channel that directed melt waters into the Gaylordsville basin.

69.6 Bear right (northwest) at intersection of Routes 55 and 39.

71.1 Descend proximal slope of outwash head. Materials grade from this ice-marginal position to the threshold near the intersection of Routes 55 and 39.

72.6 Turn right (north) and cross Tenmile River.

72.8 Turn right (east).

73.2 Outwash materials along this reach of Tenmile River correlate with materials in the Gaylordsville basin (Kelley, 1975).

74.2 Turn right (east) at Dogtail Corners.

75.7 Aerial photographs and ground traverses indicate that abandoned channels, comparable in size to the modern Housatonic River, scar the terrain to the right. They appear to record the "wanderings" of the Late Wisconsin and Holocene Housatonic River as it sought to re-assume some preglacial channel from which it had been diverted. The modern river course is not graded through this reach of the Housatonic Valley. There is no evidence to suggest the location of the ancestral channel.

This is the last stop of the trip.

75.8 Turn left (northeast) onto Route 7. Proceed northward 46 miles.

121.8 Monument Mountain School.

Basement-Cover Rock Relationships in the Pittsfield East Quadrangle, Massachusetts *

by

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Introduction

The purpose of this trip is to demonstrate the stratigraphy within the Dalton Formation and the unconformable relationship that exists between the Dalton-Cheshire clastic sequence and the underlying Precambrian gneisses. Both of these units are found in thrust slices that have overridden the autochthonous carbonate rocks along the western front of the Berkshire massif. Detailed mapping by Alavi (1971), Cullen (unpub. data), and N.M. Ratcliffe (unpub. data) has indicated that a series of as many as seven separate but overlapping thrust slices involving the basement gneiss are present (Fig. 1). They are structurally overlapping in an imbricate fashion to the northeast.

The Dalton-Cheshire sequence is late Precambrian (?) and Early Cambrian in age, on the basis of fossils found in the upper part of the Dalton Formation on Clarksburg Mountain at the south end of the Green Mountains (Walcott, 1891). No fossils have yet been found in the Dalton-Cheshire sequence attached to the Berkshire massif in western Massachusetts.

The lower part of the clastic sequence, the Dalton Formation, includes quartz pebble and cobble conglomerate, biotite-quartz-plagioclase schist and gneiss, biotite-muscovite schist, meta-arkosic sandstones and metaquartzite. The Dalton is gradational laterally and vertically into the clean metaquartzite sequence of the Cheshire Quartzite and it varies in thickness from 200 to 2,000 feet. The Cheshire, in turn, grades upward into the shelf-carbonate sequence of the Stockbridge Formation. The Dalton-Cheshire sequence interfingers eastward with schistose rocks of the Hoosac Formation which crops out east of the Berkshire massif and which may be 1,200 to 10,000 feet thick (Norton, in press). The Dalton exposures at Stop 2 has just such a schistose horizon in it (p€€dbs). At this locality, however, it appears towards the top of the Dalton Formation. The interfingering between the Dalton and the Hoosac and the thickening of the Hoosac to the east suggests an eastward-deepening basin of deposition in late Precambrian and Early Cambri-

*Publication authorized by the Director, U. S. Geological Survey

an time.

Stratigraphy

Precambrian rocks

The Washington Gneiss (p€w) is named for a blue quartz-bearing graphitic gneiss and biotite gneiss exposed near Washington, Massachusetts (Emerson, 1899). The Washington is dark colored, rusty-weathering, coarsely ribbed, blue quartz-muscovite-biotite-plagioclase gneiss and muscovite-biotite schist. A white, mica-poor, garnet-plagioclase-rich, blue-quartz-bearing granulite is interlayered on a fine scale and locally is found in the place of the more common rusty-weathering variety. Sedimentary features that may be found locally in both include thin quartz pebble conglomerate beds and graphite flakes as large as 1 cm. in diameter. Biotite-rich mafic clots are observed to be retrograded from garnet. Dark-greenish-gray to punky yellowish-brown weathering actinolite, graphite-scapolite-pyrite-bearing calc-silicate rocks, or rusty-weathering diopside-calcite marble (p€wcs) are found locally within the Washington gneiss.

The biotite gneiss (p€bg) is a biotite-hornblende-plagioclase paragneiss. It is a well-layered gray and dark-gray gneiss containing minor layers of amphibolite. An example of the amphibolite horizons that occur within it can be seen cropping out at the north end of Day Mountain.

The Tyringham Gneiss (p€t) is a light-gray to pinkish-gray weathering granite to granodioritic biotite gneiss, named for exposures in the Lee-Tyringham area (Emerson, 1899). A distinctive quartz rodding is produced by the intersection of two or more oblique cleavages. The quartz is uniformly distributed throughout the rock rather than being concentrated in layers. Field relations support the view that the Tyringham was intrusive into the older paragneisses during late Precambrian time. The emplacement of the Tyringham postdates a deformation that produced a foliation in the older rocks, and predates a deformational episode prior to the deposition of the Dalton Formation (upper Precambrian? and Lower Cambrian), (Ratcliffe and Zartman, in press).

Age of the Precambrian rocks

The Precambrian gneisses of the western front of the Berkshire Highlands have been dated at 1,040-1,080 m.y. on the basis of U-Pb isotopic data from two distinct lithologies in the Precambrian gneissic complex (Ratcliffe and Zartman, in press) (see Trip B-6). This information, in combination with the geometric evidence for an angular unconformity that is so clearly shown at Day Mountain, requires an intense pre-Dalton dynamother-

mal event.

Paleozoic rocks

The Dalton Formation (p€€d) in western Massachusetts is characterized by striking lateral and vertical variations in lithology (Fig. 2). The basal conglomerate (p€€dc) that rests unconformably on the Precambrian gneiss is composed of white fine-grained quartz, rare blue quartz, and gneiss pebbles and cobbles set in an arkosic matrix. Tourmaline and magnetite are local accessory minerals. The basal conglomerate grades into a greenish-gray muscovitic arkosic quartzite (p€€dg) containing conglomerate lenses locally. Above this is a vitreous quartzite and quartz pebble conglomerate (p€€dqc). The vitreous quartzite is strikingly similar to the Cheshire Quartzite that overlies the Dalton Formation, as seen at Stop 2. The uppermost conglomerate is a massive, white-weathering silica-cemented quartz pebble conglomerate. The vitreous quartzite is gradational laterally and vertically into a tan-weathering muscovitic-feldspathic quartzite and flaggy quartzite beds (p€€dq). A dark-gray to black biotite-quartz schist and schistose quartzite horizon(p€€dbs) appears towards the top of the feldspathic quartzite that thickens in an easterly direction. The black schistose facies closely resembles rocks mapped within the Hoosac Formation (Norton, in press) and probably represents a deep-water facies of the Dalton.

The Cheshire Quartzite (€c) is a massive, white-pinkish-tan-weathering, vitreous metaquartzite. It is laterally equivalent and interbedded with the flaggy quartzite beds of the Dalton Formation. The Cheshire Quartzite passes upward into the relatively quartz-free shelf-carbonate sequence of the Stockbridge Formation (O€s).

Unconformity

At many localities where the contact of the Dalton with the older gneisses has been observed, from Monterey north to Dalton, a peculiar, extremely muscovite-rich zone is found at the top of the gneiss immediately beneath the basal beds of the Dalton. Gneissic structures may be traced from the gneiss upwards into the muscovite-rich zone, and these relict gneissic structures are unconformably overlain by the quartz pebble conglomerate of the Dalton (p€€dc). The angular unconformity (Alavi, 1971) reaches a maximum of 90° in dip.

The localized zone of aluminous-feldspathic-magnetite-rich rock as much as several meters thick at the top of the basement gneiss may be upper Precambrian or Lower Cambrian metasaprolite. The mineral assemblage and the preservation of the Precambrian gneissic banding are consistent with this interpretation. The occurrence of this residual material at the top of the Precam-

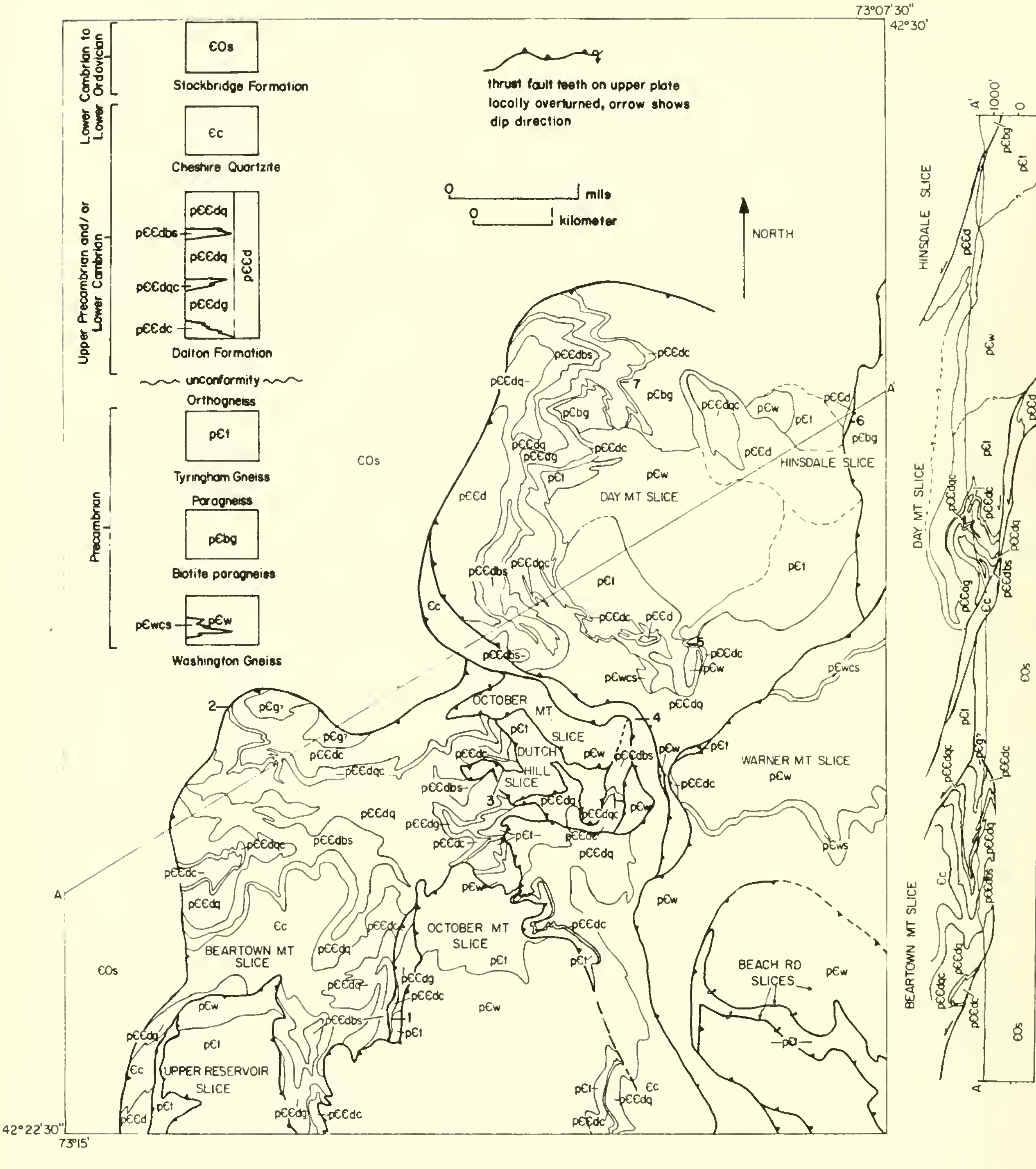
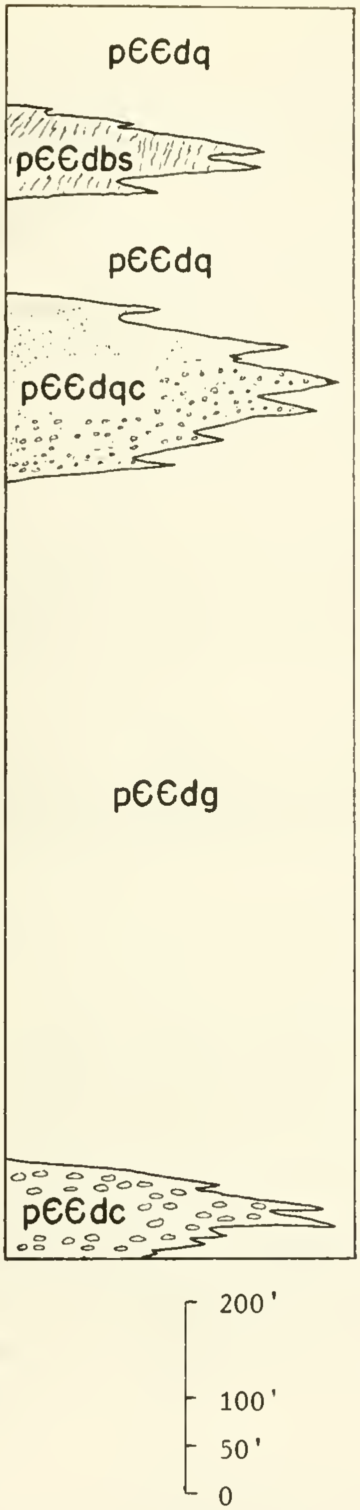


Figure 1. Preliminary geologic map of the Pittsfield East quadrangle, Massachusetts. Cross-section A-A′ (on the facing page with Fig. 2) is approximate. Data from Alavi (1971), Cullen (unpub. data), and Ratcliffe (U.S.G.S. unpub. data). Stop locations numbered 1 - 7.



pЄЄdq 300' 150' Feldspathic quartzite and flaggy quartzite beds 10 cm to 1m thick, interbedded with (pCCdbs) biotite, quartz schist and schistose quartzite.

pЄЄdqc 80'-200' Vitreous quartzite and quartz pebble conglomerate.

pЄЄdg 100'-750' Greenish-gray muscovitic, arkosic quartzite with conglomerate lenses.

pЄЄdc 35'-100' Basal quartz pebble and cobble conglomerate with arkosic matrix.

Figure 2. Columnar section demonstrating the relative thicknesses of the different facies in the Dalton Formation, after Alavi (1971).

brian terrane sheds some light on the character of the paleo-erosional surface. The surface probably had low relief and underwent prolonged exposure in a temperate climate, during which alkalis and alkaline earths were leached, thus enriching the saprolite in alumina, silica, iron, and combined water.

This metasaprolite is comparable to a similar horizon at the top of Precambrian rocks reported from the Southern Appalachians (Rankin, 1967).

The muscovitic lower part of the Dalton (p€€dg) probably derived much of its potassium and aluminum locally from this saprolite horizon.

Nature of Precambrian deformation and metamorphism

Complex structures within the basement gneiss suggest the possibility of multiple deformations in the Precambrian before the deposition of the unconformable cover rocks in the early Paleozoic (Alavi, 1971). The Precambrian gneisses locally possess a mineralogy consistent with sillimanite grade of metamorphism in the Grenville orogeny.

The Precambrian foliations, lineations, minor fold axes, and several major folds trend northeast to east-northeast. The structural trends in the Paleozoic rocks generally strike northwest and dip variably to the southwest (Alavi, 1971; Ratcliffe and Harwood, in press).

Deformation in the Paleozoic

Rocks of the Berkshire massif have been thrust westward across the autochthonous shelf sequence (see Trip B-6, fig. 1 for a regional map). In the Pittsfield East quadrangle, seven slices are shown(Fig. 1 this paper) and further identified in the explanation of figure 1 (Trip B-6). Locally along the soles of the overthrusts, intense zones of cataclasis and recrystallization are found associated with large recumbent folds (Ratcliffe and Harwood, in press). Section A-A′ (Fig. 1) shows the structural relations as presently conceived. Multiple deformations in the Paleozoic postdated the thrusting and will not be discussed in detail here (see Trips B-2, B-6).

Acknowledgements

I would like to thank Professor N.M. Ratcliffe for his assistance in preparing this manuscript, both in the field and with editorial remarks.

ROAD LOG

Stops for this trip will be within the Pittsfield East, Massachusetts quadrangle. Trip log begins at the Monument Mountain Regional High School parking lot.

Mile	
00.0	Depart Monument Mountain Regional High School at 8:30 A.M. sharp! Turn right onto Rt. 7, driving north.
02.7	Entering the town of Stockbridge, Red Lion Restaurant on the right. Turn right at "T" intersection (Rts. 7 & 20). Follow signs for Rt. 7 out of Stockbridge.
03.0	Turn left following signs for Rt. 7; continue north.
07.4	Rts. 7 & 7-A branch here; stay on Rt. 7 (straight).
08.3	Stop light at intersection of Rts. 7 & 20; turn left continuing north on Rt. 7.
12.0	Prepare for sharp right turn at the bottom of hill.
12.4	Turn right onto New Lenox Rd., condominium project on the right (The Colony). Proceed east on New Lenox Rd.
13.2	Stop sign at intersection of New Lenox Rd. and East St. Continue east on New Lenox Rd. (straight). The hills rising above the valley we are crossing are composed of Precambrian gneisses and overlying lower Paleozoic cover rocks (p€€d and €c). This is part of the Beartown Mountain slice that has been thrust above the marble (O€s) that underlies the valley. The fault scarp can be seen extending to the south.
14.3	Intersection of New Lenox Rd. and East New Lenox Rd. to the left and October Mountain Rd. to the right. A gray Federal house is on the left and the New Lenox Cemetery is ahead on the right. Proceed through the intersection continuing east.
15.1	Pumping station on the right. The rocks below in the brook are feldspathic quartzites of the Dalton Formation. They are stratigraphically above the black schist facies of the Dalton. The feldspathic quartzites are in turn overlain by the vitreous quartzite of the Cheshire Quartzite. Dewey Hill to the south is capped by Precambrian gneisses (Tyringham and Washington) in a south-dipping thrust sheet that discordantly overlies the lower Paleozoic rocks in this locality.

16.2 Bear left at the branch in the road, over the brook and up.

16.9 Farnham Reservoir on the right. If the water is low enough, Cheshire quartzite (Ꞓc) can be seen exposed on the north shore.

17.4 Stop 1. Farnham Reservoir section
This stop demonstrates the manner in which the unconformity is usually seen in the field. The Precambrian gneiss exposed in the small brook dissecting the slope on the left side of the road is of Tyringham (pꞒt) lithology. The gneissic banding strikes east and dips steeply south. The conglomerate facies of the Dalton (pꞒꞒdc) is exposed on the slopes to the right of the road, and in the road itself. The pebbles in the conglomerate are white and black quartz and an occasional gneiss or amphibolite clast. These clasts are set in an arkosic matrix (K-feldspar). Magnetite and tourmaline are minor constituents. The latter seems to parallel the long axis of the pebbles at many localities. The contact between the Precambrian gneiss and basal Cambrian conglomerate is not physically exposed at this locality. There is, however, a distinctive zone of muscovite (weathers apple-green) -rich rock between the two. The muscovite is aligned in a foliation plane common to both units. This zone of strange(?) rock is quite variable in thickness, but is very often associated with the unconformity. The presence and significance of this rock has been the subject of much speculation among those who have encountered it in the field. We hope that if these relations are demonstrated in the field, some interesting ideas may evolve. This particular exposure is on a thrust sliver caught between the Beartown Mountain slice below and the overlying October Mountain slice (Fig.1).

Return to cars and drive slowly south along the road.

17.5 The Tyringham gneiss exposed at the bend in the road is associated with the leading edge of the October Mountain slice. The gneissic layering is nearly vertical, whereas a prominent cataclastic foliation dips east into the hill.

17.9 Turn around, using the short lane that intersects the road from the right; be careful if it is wet. Retrace the route back down the hill to the intersection with the gray Federal house on the right.

21.5 Turn right onto East New Lenox Rd. and drive north. The

hills to the west are rocks of the Walloomsac Formation associated with the Everett slice of the Taconic allochthon.

23.5 Stop 2. Sykes Brook-Power Line traverse (Figs. 1 & 3) A church parking lot is on the left side of the road. Hopefully we can obtain permission to use it. Failing that, pull off to the side of the road, trying not to block access to the fields or sheds by the roadside. The purpose of this traverse is to demonstrate the variation in lithology within the Dalton Formation. Two trails intersect the road at this point, we will walk east on the lower trail closest to the brook. Hike up the trail to the shed on the left side of the trail, turn left and walk down the slope to the brook.

a. The following discussion will be treated as a continuous traverse up the brook to the power line. The Dalton exposed in the brook grades from a muscovite-feldspar-quartz schist to a feldspathic quartzite. Abundant recumbent folds that have an axial planar cleavage are seen in the feldspathic quartzite. A well-developed lineation has been caused by the intersection of bedding and cleavage. The axial planar foliation is seen to dip first in one direction, then another. This variation in attitude is due to post recumbent folding. Upstream, exposures of massive vitreous quartzite are seen on the slopes to the left. These are similar to Cheshire (Ꞓc) lithology, but the stratigraphic position is lower than the younger Cheshire. Below this, isolated exposures of the stretched-pebble conglomerate are seen (pꞒꞒdqc).

Return to the trail. At this juncture, participants have the option of returning to the cars via this trail or continuing the traverse along the power line ahead. This part of the traverse will be somewhat strenuous because of thick underbrush and rough terrane.

b. This part of the discussion deals with the rocks seen on the second leg of this traverse; along the power line and back to the cars via Sykes Mountain Rd. Up to this point we have been going down section. Now, we are on the southeast limb of an overturned syncline, and the traverse along the power line will take us back up section. Southwest along the power line, very good exposures of the stretched-pebble conglomerate with arkosic matrix are found. Bedding and cleavage relations are particularly well-developed; this feature and crossbeds, when seen, indicate the tops are to the east or southeast. Farther along this exposure, the conglomerate gives way to the vitreous quartzite, which in turn grades into schist and

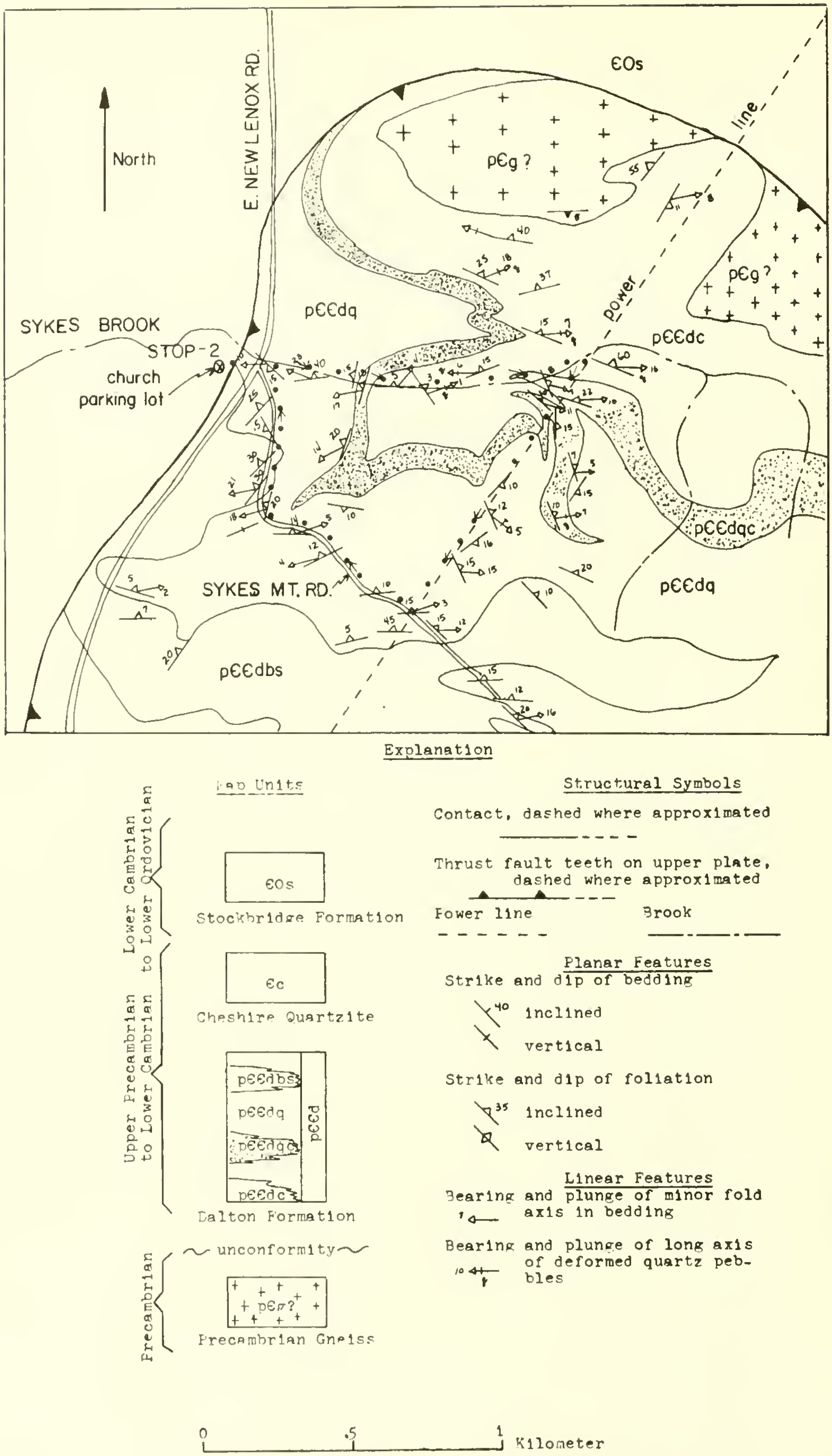


Figure 3. Detailed geologic map of Stop 2.,(. . . . indicates the path of the traverse to be taken).

interbedded quartzite at the intersection with Sykes Mountain Rd. Returning to the vehicles via this road we will walk into the black schist facies of the Dalton (p€€dbs). This unit is somewhat similar to Hoosac rocks mapped to the east as a basal Cambrian unit; the black schist here is near the top of the Dalton Formation. Complexly refolded features can be seen in exposures in the trail and on the slopes to the left. The foliation is folded into a synclinal structure. We are now walking down section on the way back to the cars.

Reassemble at the cars and proceed north on East New Lenox Rd.

24.6 Turn right at "T" intersection. This is William St. (may not be marked). Drive east on William St.; Mt. Greylock can be seen on the left.

26.1 Intersection of Washington Mountain Rd. and William St.; bear right. The road is now Washington Mountain Rd.; the Dalton-Pittsfield town line was also at the intersection.

26.3 Bear right at "Y" intersection, proceed southeast, still on Washington Mountain Rd.

27.9 Stop 3. Hathaway Brook section (Figs. 1 & 4)
Pull off to the right where the shoulder widens. Traffic is sometimes brisk along here, be alert! If you have topographic maps, we are at the point where the 1650' contour intersects the road. The first outcrop is back down the road about 600 feet. At this stop, we will look at the augen gneiss in the sole of the Dutch Hill slice (Fig. 1) which is exposed on the road. In the brook below the road, Precambrian gneiss can be seen in thrust contact above rocks of the Dalton Formation. The rock exposed in the roadcut is Tyringham gneiss that contains quartz and feldspar augen. It is a mylonite gneiss formed along the sole of the thrust slice that rests above the Beartown Mountain slice and below the Dutch Hill slice. This rock is figured in Ratcliffe and Harwood (in press, fig. 15). The blastomylonitic foliation consists of crushed and granulated feldspar with biotite aligned in it; it is oriented N.75W. and dips gently to the north. In the brook below the road, the north-dipping thrust contact is exposed. This exposure could be confusing because it is the basal conglomerate of the Dalton that is in contact with the Precambrian gneiss. It could be interpreted as an overturned sequence, but the adjacent field relations preclude this possibility (Fig. 4).

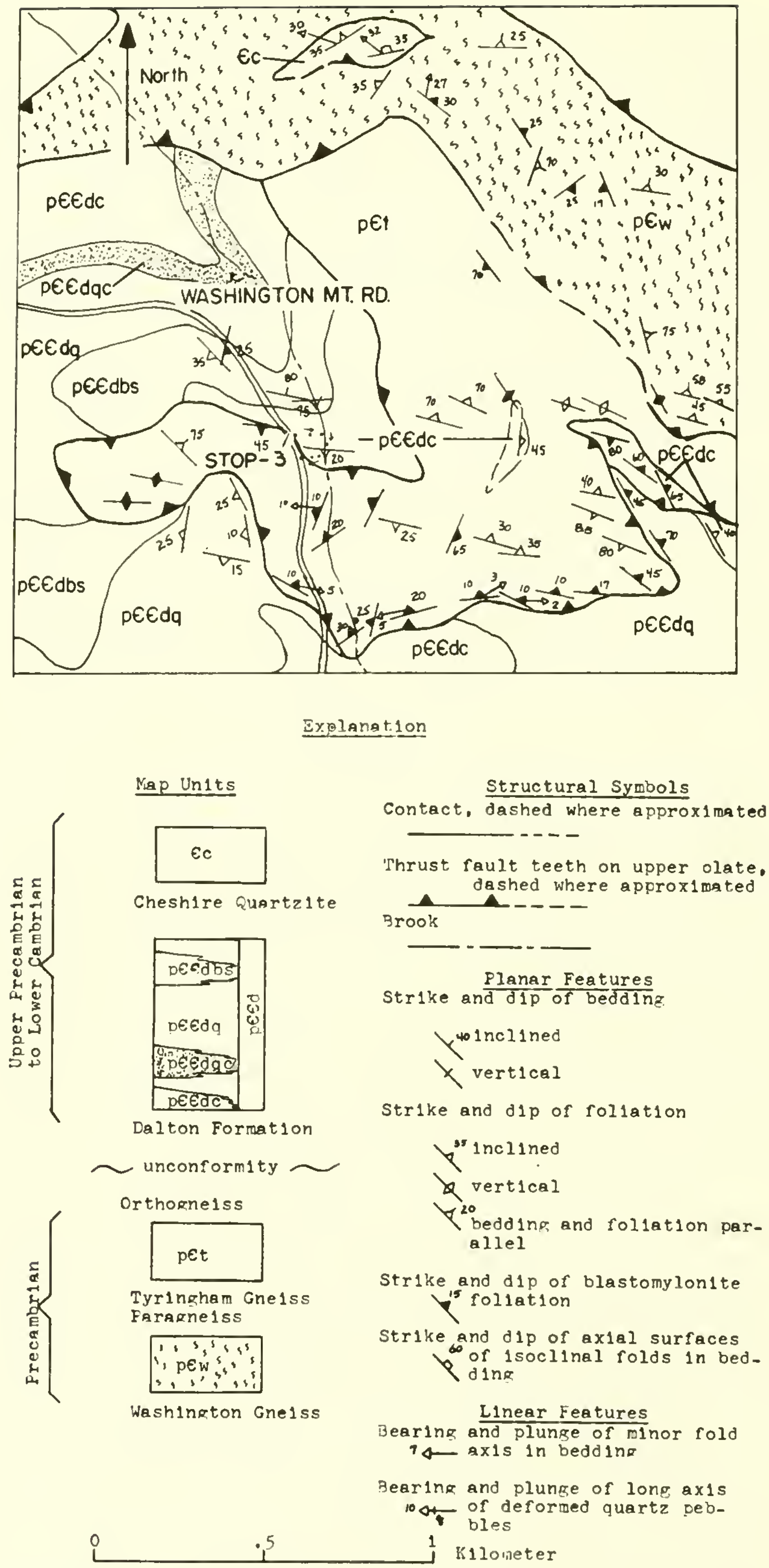


Figure 4. Detailed geologic map of Stop 3., (. . . . indicates the path of the traverse to be taken).

Return to the cars and turn around; proceed back down Washington Mountain Rd.

29.6 At the intersection of Washington Mountain Rd. and Kirchner Rd. (yellow house on the right), turn right onto Kirchner Rd. and drive east.

31.6 Stop 4. Upper Sackett Reservoir - LUNCH STOP
Pull off to the right where the access road to the reservoir intersects Kirchner Rd. At this stop, those who wish may walk up the road to the reservoir, eat lunch, and enjoy the scenery. Others may wish to look at some of the geology while they eat. There is a fine exposure of a hematite-cemented breccia in Cheshire quartzite on the slopes below the road. The Cheshire at this locality is on a sliver caught between two larger thrust slices: the Day Mountain slice is above to the north and a sliver of Precambrian gneiss (pЄw) is exposed in the dam spillway, structurally below.

After lunch, return to the cars and continue ESE on Kirchner Rd.-Pittsfield Rd.-Blotz Rd. The name changes on the map, but it is the same road.

32.5 Crossing the contact of pЄЄd and pCw of the Warner Hill slice above the Day Mountain Slice.

33.1 Turn left onto a partially concealed logging road (this route will be taken only if we have all-wheel drive vehicles available). Drive the length of the road until it terminates on a knoll above the power line. If we cannot use this road, an alternate route is described later, as it is on the way to the next stop. It requires a vigorous hike, however, and to save time, the first alternative is desired. To avoid confusion, the road log will not include the mileage along this logging road and will resume at the intersection of Blotz Rd. on the way out.

Stop 5. Belmont Reservoir section
At this stop, we will be able to observe one of the few truly fine exposures of the unconformity. Walk NNE from the knoll above the power line, across the power line following a logging trail. Descend into the swale on the other side of the power line. The unconformity is exposed to the northeast in this brook at a lower level. It is best reached by keeping to the high ground on the left. As you approach, note the good gneissic qualities of the outcrops. At the unconformity, the gneissic banding is seen to be truncated by the overlying conglomerate.

Both the pЄЄdc and the pЄt demonstrate the common orientation of the apple-green-weathering muscovite in a late foliation. Note the flattened pebbles in the pЄЄdc. The clasts are white and black quartz and an occasional gneiss or amphibolite pebble. Farther downstream, the Dalton becomes more schistose.

Return to the cars and turn around and return to Blotz Rd., where the log will resume. Turn left on Blotz Rd. and drive southeast.

33.4 Turn left onto Plunkett Rd. and drive north past the Hinsdale town dump on the left and the Plunkett Reservoir on the right. The road surface improves here. Slow down when passing Camp Emerson (on the left); the alternate turn-off for the Belmont Reservoir section is coming up.

35.5 Alternate approach to Belmont Reservoir: Turn left onto the small unmarked access road that leads to the reservoir. It may be chained; if we have to use it, arrangements will be made to have it unlocked. Follow the access road until it terminates at Belmont Reservoir. Follow the brook that empties into the southwest corner of the reservoir up towards its source. This traverse will be down section, beginning in the schistose facies of the Dalton and down into the basal conglomerate at the unconformity farther upstream.

Continuing the log from the alternate turn-off: Continue north on Plunkett Rd. to unmarked intersection (next one).

35.6 Turn right on unmarked road, follow it around to the left where it parallels the R.R. tracks.

35.9 Turn right over the bridge that spans the R.R. tracks.

36.0 Turn left at the stop sign ("T" intersection).

36.1 Turn left again at the next stop sign, onto Rt. 8, and drive north.

37.1 <u>Stop 6</u>. Hinsdale R.R. section
Pull off the road to the right just beyond the Dalton town line. A dirt road that intersects Rt. 8 from the right may afford some space. Cross the road and enter the R.R. cut. Trains do use these tracks -- watch out!

The purpose of visiting this locality is to observe the

effects of thrust faulting in the Precambrian gneiss and the Dalton, both of which are exposed at this locality. We are at the contact of the thrust that marks the sole of the Hinsdale slice. At the south end of this exposure, the rocks are a biotite-quartz-plagioclase paragneiss. These rocks are complexly folded and sheared. Preliminary slip-line analysis using the orientation of drag-fold axes in the plane of shearing (after Hansen, 1971) has indicated an ENE direction of transport. The presence of chlorite in these gneisses indicates a retrogressive metamorphism of an earlier (Precambrian) high-grade metamorphic facies. At the north end of the R.R. cut, possible Dalton is exposed. A pink rock, rich in muscovite is caught in the sole of the Hinsdale thrust. The Hinsdale slice is one of the tectonically highest slices in the quadrangle. The cataclastic foliation along this section has a northeast strike and dips south.

Return to the cars and proceed north along Rt. 8.

38.4 Turn left by an abandoned stone mill (on the left) onto East Housatonic St.

38.6 Turn left onto East Rd. (V.F.W. on the right); drive south.

39.2 At stop sign, turn right onto Rt. 8; drive south towards Hinsdale.

39.9 Bear right at the Shell station (on the right).

40.0 Take first right and cross the R.R. bridge.

40.1 Right again after crossing bridge; follow the road around to the left. Leave Hinsdale, driving northwest on Robinson Rd.-Grange Hall Rd. (probably not marked).

42.8 Stop 7. Day Mountain section (Figs. 1 & 5)
Park the cars on the shoulders to the right. This is the section referred to in great detail by Emerson (1899). It demonstrates admirably the unconformable relationship between the Precambrian gneiss and the Dalton Formation. The traverse begins in the brook below the road, where Precambrian gneiss (Hinsdale Gneiss of Emerson; p€bg this paper) is nearly vertical, striking almost due east. These gneisses have been dated at 1 b.y. (Ratcliffe and Zartman, in press). The presence of epidote in fractures that cut across and locally offset the gneissic banding is suggestive of a retrograde metamorphic event in the early

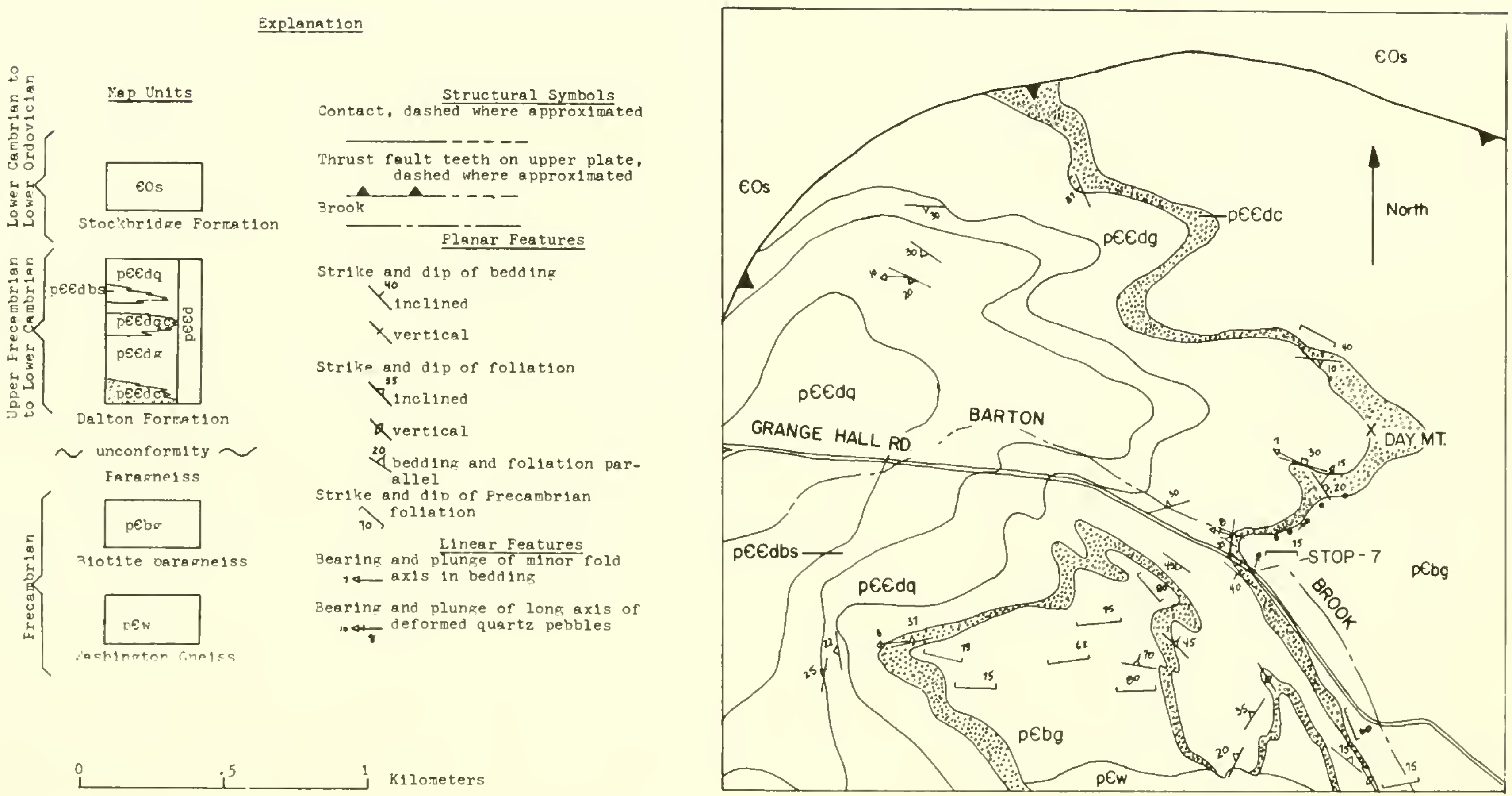


Figure 5. Detailed geologic map of Stop 7., (. . . indicates the path of the traverse to be taken).

Paleozoic. Up the slope towards the summit of Day Mountain, on the north side of the brook, the gneiss gradually changes in character, similar to what we have seen at other localities. As we approach the unconformity, the rocks become richer in muscovite. At the contact, the unconformity is dramatically exposed. The gneissic banding, preserved as a relict structure in the metasaprolite, is truncated by the basal conglomerate of the overlying Dalton Formation. On the basis of the relationships seen here, it is not surprising that contacts between Precambrian and lower Paleozoic rocks are confusing and seemingly concordant at many localities that are less well exposed (Stop 1) than here.

Return to the cars and drive northwest on Grange Hall Rd.

43.7 Turn left at the Grange Hall onto South St.; drive south.

44.6 Turn left again onto Division Rd. and drive south.

45.3 Turn right onto Washington Mountain Rd.-William St.; drive west. Stay left at "Y" intersection (William St. and Elm St.).

45.6 Turn left at traffic light onto Holmes Rd. and drive south.

51.9 At traffic light turn left onto Rt. 7; return to Great Barrington via Rt. 7 south.

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Trip B-10

GENERAL GEOLOGY OF THE STOCKBRIDGE VALLEY MARBLE BELT

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Introduction

The bedrock geology of the Stockbridge Valley marble belt consists of rock units ranging in age from Precambrian to Middle or Upper(?) Ordovician. Mapping in several quadrangles in this general area (including the Stockbridge and Ashley Falls quadrangles which are visited in this trip) reveals a complex history of thrusting, folding, and metamorphism. The clearest manner in which to view the geologic relationships and map patterns is to divide the various rock units into lithotectonic sequences as proposed by Ratcliffe (1975). This approach allows the bedrock to be divided into three sequences, classed in relative terms as autochthonous, parautochthonous, and allochthonous. In general, the Walloomsac and Stockbridge Formations are restricted to the autochthonous sequence; the Everett Formation comprises the allochthonous sequence; and the parautochthonous sequence contains Cheshire Quartzite, Dalton Formation, and Precambrian gneisses. Folding and metamorphism both predate and postdate periods of faulting, thus producing folded thrusts and a pattern of numerous disconnected klippe.

Stratigraphy

Rock units examined during this trip consist of the Lower Cambrian to Lower Ordovician carbonates of the Stockbridge Formation, the Middle Ordovician Walloomsac Formation, and the Cambrian(?) and Lower Cambrian Everett Formation. The seven lithic subdivisions of the Stockbridge Formation proposed by Zen and Hartshorn (1966) are exposed in the Stockbridge and Ashley Falls quadrangles. Five of these units, plus a new quartzitic unit, will be visited during this trip.

Detailed descriptions of the carbonate units and other formations are included at the appropriate position in the road log.

Structure and Metamorphism

The Paleozoic rocks in the Stockbridge and Ashley Falls quadrangles have experienced four episodes of folding, three of which were accompanied by some degree of metamorphic activity. Refolded fold patterns, folded foliations, cleavages, and lineations of various categories clearly document the interrelationships of folding, faulting, and metamorphism. In addition, there is firm evidence that a major unconformity exists beneath the Walloomsac Formation as it rests on all units of the Stockbridge Formation in the Ashley Falls quad-

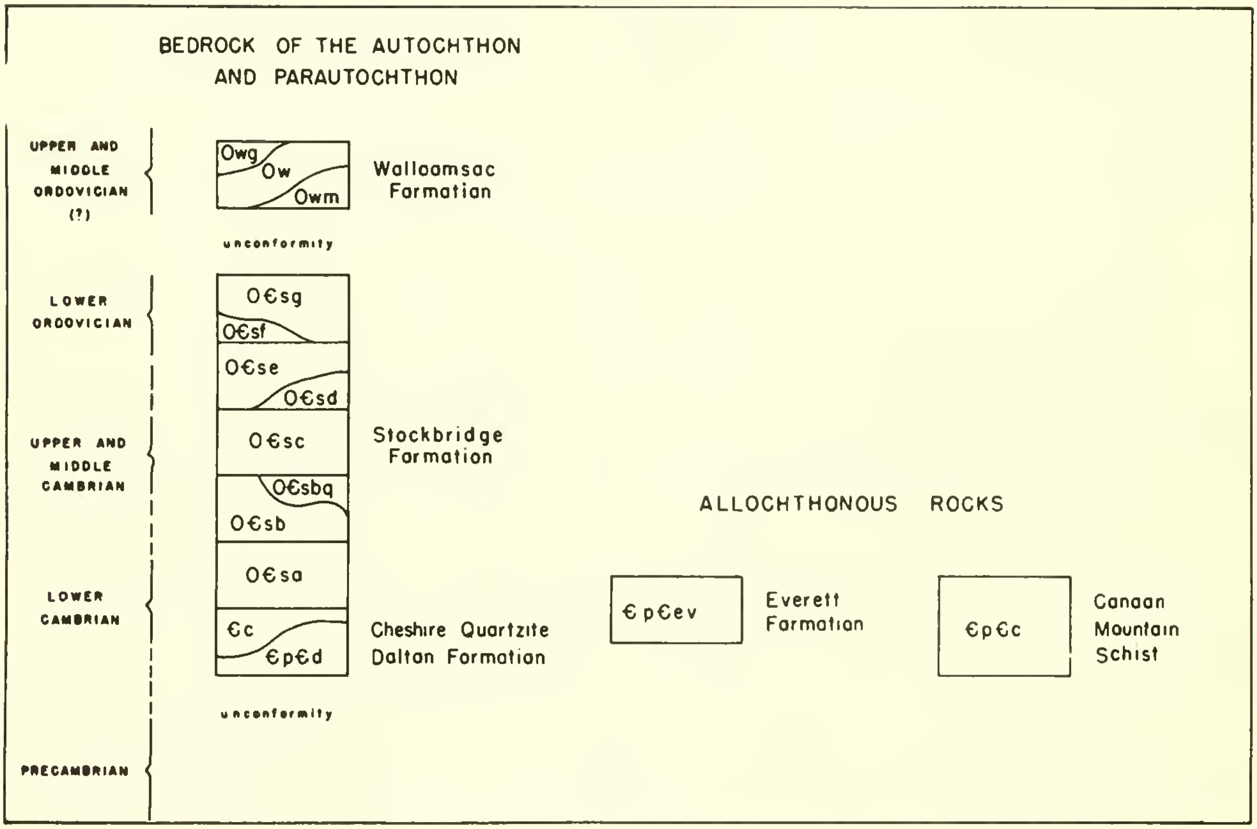

Figure 1. Stratigraphic sequence of rock units discussed in this field guide.

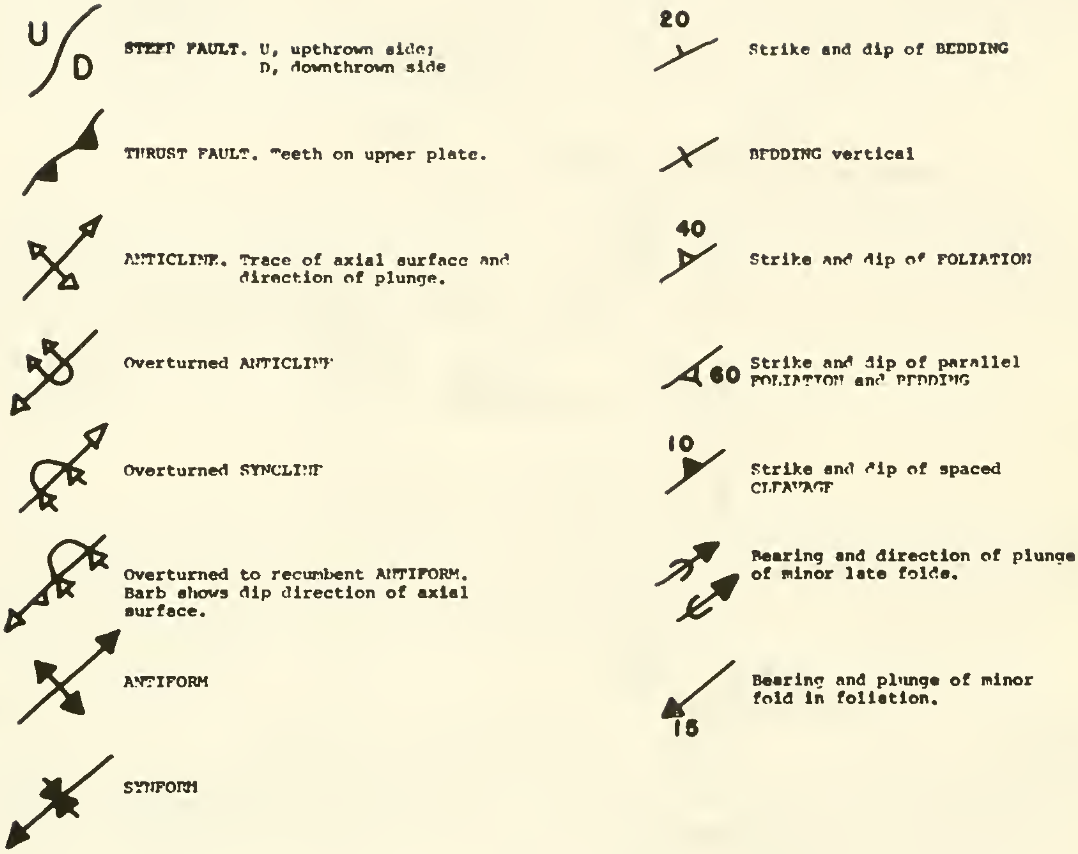

Figure 2. Explanation of structural symbols used in this guide.

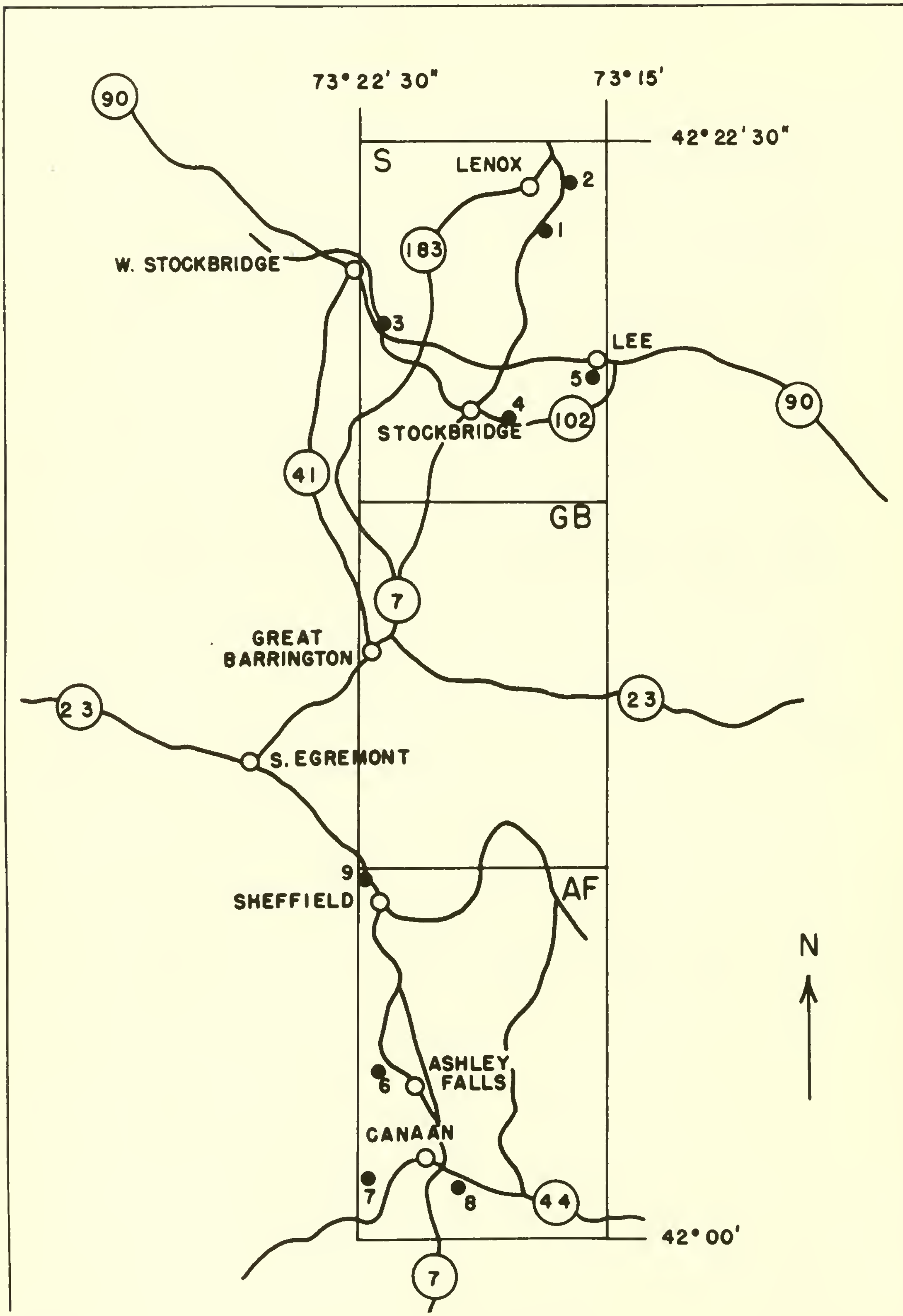

Figure 3. Map showing stop locations and numbers.
S = Stockbridge quadrangle; GB = Great Barrington quadrangle;
AF = Ashley Falls Quadrangle.

rangle. Much of the evidence for the inferred tectonic history of the area will be developed during the course of this trip. Metamorphic grade increases from the northwestern corner of the Stockbridge quadrangle where the garnet isograd is present to the southeastern third of the Ashley Falls quadrangle where the sillimanite-muscovite isograd is exposed.

Acknowledgements

Much of the material in this guide for trip B-10 is based on the field geology of Nicholas Ratcliffe. I wish to express my appreciation for his willingness to share his data for one of my personal research projects and for his encouragement and example while mapping in the Ashley Falls quadrangle.

Road Log for Trip B-10

Assemble at 8:30 a.m. at the Monument Mountain Regional High School parking lot. Bring lunches as no provision for grocery purchases will be made. Stops will be in the Stockbridge, Mass. and Ashley Falls, Mass.–Conn. quadrangle. General trip route and stop locations are shown in Figure 3.

<u>Mileage</u>

0.0	Parking lot, Monument Mountain Regional High School.
0.1	Turn right upon leaving parking lot and proceed north on Rt. 7.
0.7	Cross Stockbridge town line.
1.8	Roadside outcrop on left is in Unit C, Stockbridge Formation.
2.6	Junction of Rt. 7 and Rt. 102. Turn right onto 7-102. Pass through the center of the town of Stockbridge.
3.0	Rts. 7 and 102 split. Keep to left and continue north on Rt. 7.
4.0	Outcrop on left in Unit C.
4.2	Massachusetts Turnpike passes overhead.
5.2	Cross Lee town line. The hills immediately to the left (west) are underlain by Dalton Formation which has been thrust over tightly folded Stockbridge and Walloomsac rocks (parautochthonous sequence over autochthonous sequence).
6.5	Entrance to Fox Hollow School on right.
6.8	Outcrops to left are in Unit C. Immediately west of these outcrops Units D and E are exposed in an inverted anticline that has been refolded by a northerly plunging antiform.
7.2	Intersection of Rts. 7A and 7. Continue on Rt. 7.
8.1	Intersection of Rts. 7 and 20. Continue on 7.
8.3	<u>STOP 1</u>. Pull off on right shoulder. Exercise CAUTION due to heavy traffic. Refer to Figure 4 for general structural relationships. These roadcuts are in Unit D of the Stockbridge Formation. This unit consists of quartzitic calcite marble with subordinate beds of gray micaceous dolostone and sandy dolostone. Occasionally present are thin, rusty-weathering schist partings. At this stop Unit D is deformed into a recumbent fold which deforms schistosity and is warped by later, more open folds. Petrofabric analysis shows evidence of both periods of deformation. Unit C is exposed just to the south along the highway.

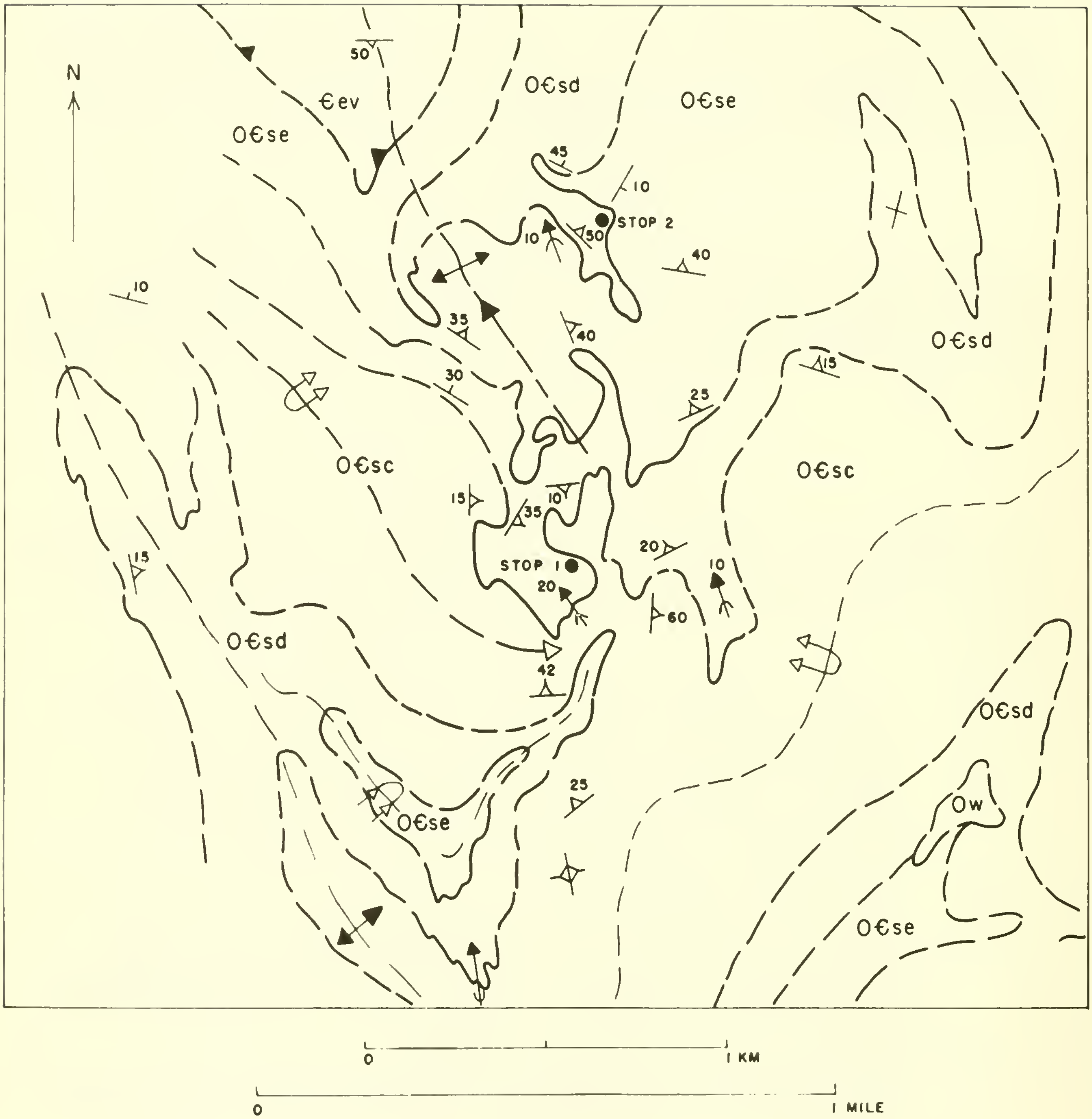

Figure 4. Generalized geologic map of the area surrounding Stops 1 and 2 (after Ratcliffe, 1975).

Mileage

8.8 STOP 2. Intersection of Rt. 7 and Housatonic St. Pull over on right shoulder. We will examine several outcrops along this busy highway. Please be CAREFUL. The map pattern here is dominated by an early period of folding which produced overturned and inverted folds which have been refolded along north to northwest trending axes (see Fig. 4). Units D and E are involved here. Unit E is dominantly a white and massively bedded, coarsely crystalline calcite marble with interbeds of blue-gray calcite marble. Orientations of minor folds here and at Stop 1 reflect the orientation and character of the 2 folding episodes.

8.9 Turn left onto Housatonic Road and proceed west.

9.7 T-junction. Turn left.

9.9 Turn right onto Rt. 7A-183.

9.95 Bear to left on Rt. 183 and proceed down the hill.

11.4 Main entrance to Tanglewood.

11.5 Stockbridge town line.

13.4 Outcrops of Unit B on left. Here Unit B is completely surrounded by Unit C and represents an inverted anticline that has subsequently been refolded by a synform plunging to the southwest. The high hills (West Stockbridge Mountain) to the right (west) are underlain by the allochthonous Everett Formation.

14.2 Bear right on Rt. 183. Pass through the village of Interlaken.

15.0 Outcrop of Unit C.

15.2 Massachusetts Turnpike passes overhead.

15.6 Village of Larrywaug. Junction of Rts. 183 and 102. Turn right onto Rt. 102 west.

16.5 Outcrops of Everett Formation on left.

16.7 STOP 3. Pass over Massachusetts Turnpike. Park on shoulder to right. Caution, dangerous curve. Both the main schistose member of the Walloomsac Formation and the Everett Formation are exposed here. The Walloomsac is a black to dark gray, biotite-rich, quartzose schist; silvery-gray, coarse-grained biotite schist; or light gray, punky-weathered, black fine-grained schist. The Everett Formation is essentially a greenish gray to dark green quartzose schist rich in chlorite or chloritoid, almandine garnet, and muscovite. These two formations are in thrust contact. Mapping in this general area clearly shows that the thrust contact has been deformed in a series of overturned anticlines and synclines which are refolded by southeasterly plunging structures (see Fig. 5). Numerous minor structures are also exposed at this stop.

Note: Turn around by backing into side road on right. Please take your time and exercise caution. Proceed south on Rt. 102.

17.8 Intersection of Rts. 183 and 102. Continue on 102.

19.7 Intersection Rts. 7 and 102. Town of Stockbridge. Continue east on Rt. 7-102.

20.0 Rts. 7 and 102 diverge. Keep to right on 102.

20.9 STOP 4. Exposed along this roadcut are the schistose marbles of the Middle Ordovician Walloomsac Formation. These schistose marbles are dark gray, impure, calcitic, and contain disseminated black plagioclase. The unit is often massive and lacks bedding and may contain lenses of biotite schist.

The presence of a basal metaconglomerate containing carbonate clasts at numerous localities plus the observation that these marbles are thickest where upper units of the underlying Stockbridge Formation are missing suggests that this unit was derived from erosion of the Stockbridge (Ratcliffe, 1975).

Continue east on Rt. 102.

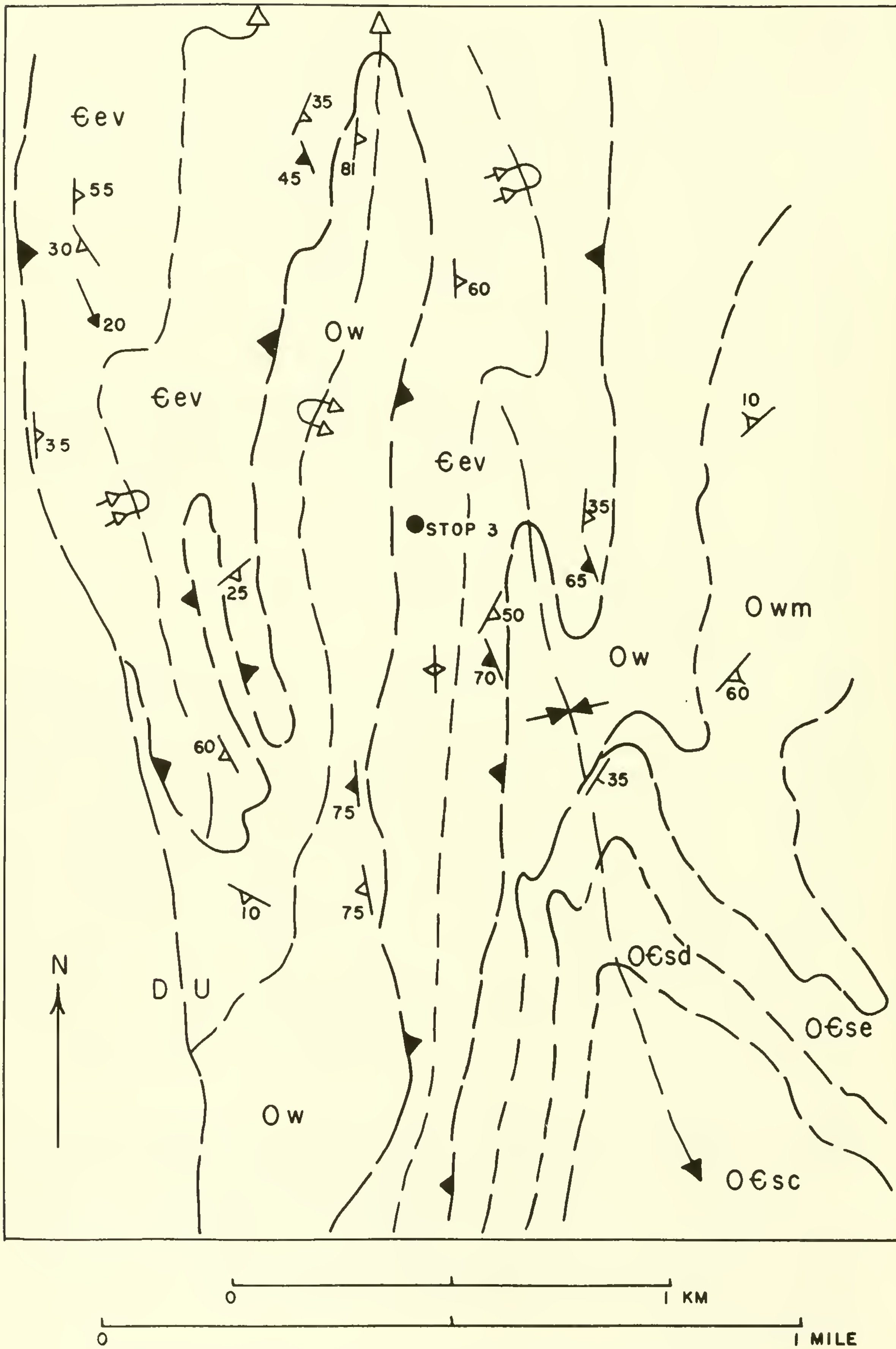

Figure 5. Generalized geologic map illustrating structural relationships surrounding Stop 3 (after Ratcliffe, 1975).

Road Log for Trip B-10 (cont'd)

Mileage

21.1 Lee town boundary.
21.7 Turn left onto Fairview Street.
23.3 Cross Willow Brook. Units A and B outcrop in stream and fields.
23.5 Pass under Massachusetts Turnpike.
24.0 Turn right onto Stockbridge Road.
24.2 Cross railroad tracks; turn right onto Marble Street.
24.5 Entrance to Lee Lime Quarry. Turn right. Temporary end to road log while in quarry.

STOP 5. Stockbridge Formation Units A and B are exposed in the Lee Lime Quarry. Unit A is a massive, coarse- to medium-grained, white dolomitic marble. This unit is quarried here for use as a high magnesian agricultural lime. Unit B is an impure quartzose, micaceous dolostone with phlogopitic and quartzose partings. It varies from gray- to tan-weathering and is massive to well-layered. It also commonly contains interbeds of rusty-weathering, schistose, feldspathic, quartzose dolostone.

The quarrying operations at this locality are controlled by the local structure (see Fig. 6). Unit B is exposed in the keel of a southeasterly plunging synform that cuts at high angles across an earlier anticline. Unit A, exposed to the northeast and southwest of Unit B, is quarried, and B is left standing as a ridge in the quarry.

Leave quarry and turn left at quarry entrance.

24.7 Turn left onto Stockbridge Road.
25.0 Turn left onto Fairview Street.
27.2 Turn right onto Rt. 102.
28.0 Pass by Stop 4.
29.2 Stockbridge town center.
29.3 Turn left on Rt. 7.
31.2 Great Barrington town line. Continue on Rt. 7.
35.4 Junction of 23W and 7S. Take 7S to right. Outcrops of Walloomsac Formation schistose marble.
36.0 Junction Rts. 7, 23, 41. Turn left; continue on Rt. 7.
36.4 Center of Great Barrington.
36.9 Continue on Rt. 7. Rts. 23 and 41 diverge to right.
39.0 Sheffield town line.
42.7 Sheffield town center.
44.5 Rt. 7A leaves Rt. 7 to right. Follow 7A toward Ashley Falls.
45.1 Turn right, cross railroad tracks, follow signs to Bartholomew's Cobble.
46.5 Y-intersection. Bear to left.
46.6 Turn right onto Weatogue Road. End of hard pavement.
46.7 STOP 6. Bartholomew's Cobble. Exposed here is Unit Bq, a distinctive quartzite that is irregularly pitted and knotted with coarse growths of quartz and bladed tremolite. This unit is discontinuous, located near the top of Unit B, and is present only in the western part of the Ashley Falls quadrangle.

Lunch stop. We will eat in the parking area, courtesy of the Trustees of Reservations. Anyone wishing entrance to the Cobble proper must pay a fee of $.50.

Continue south on Weatogue Road after lunch.

47.5 Hard surface begins. Outcrops of Walloomsac schists.
50.3 Turn right onto Twin Lakes Road.

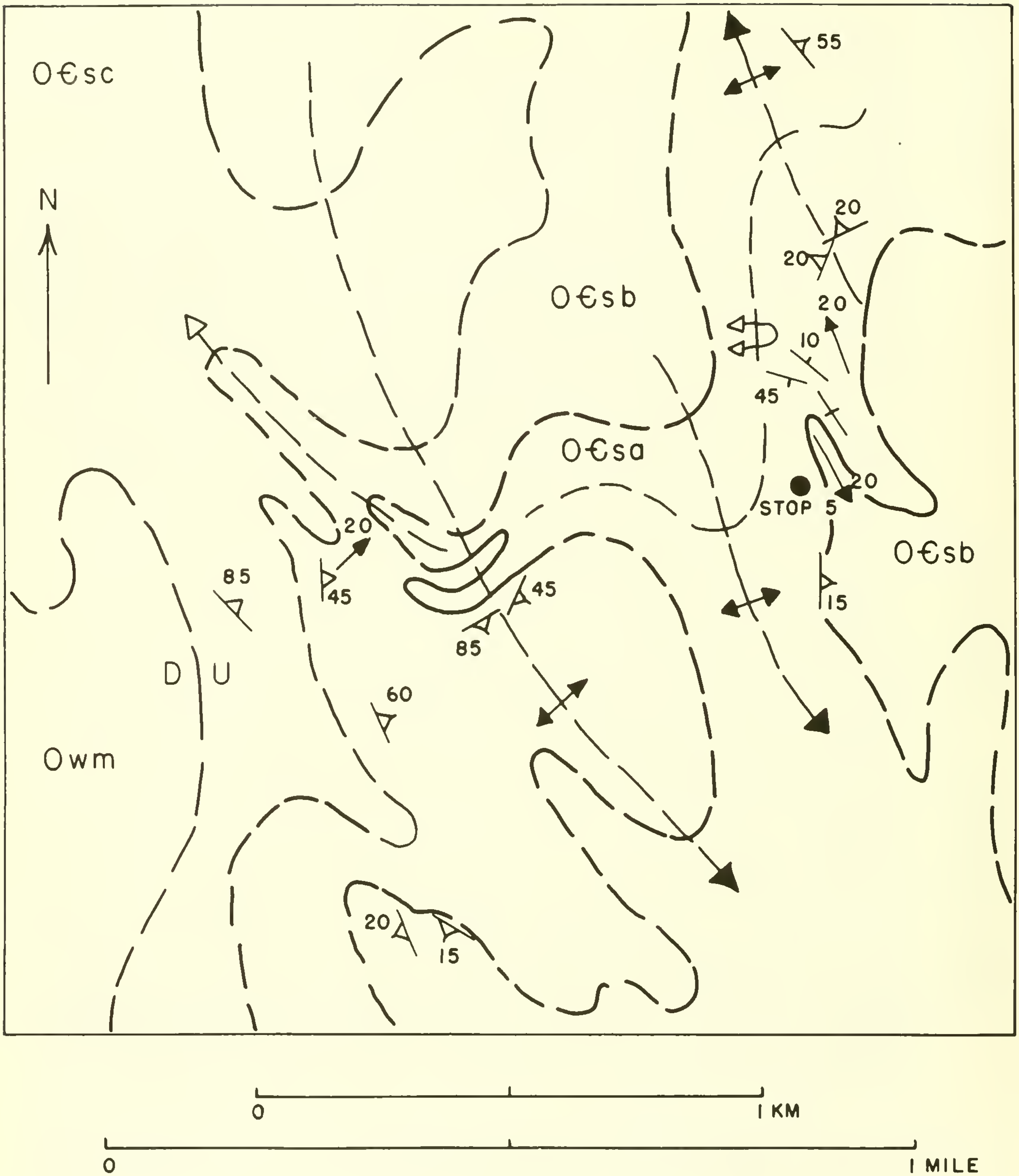

Figure 6. Generalized geologic map illustrating structural relationships at Stop 5 (after Ratcliffe, 1975).

Road Log for Trip B-10 (cont'd)

Mileage

50.7 STOP 7. Abandoned Penn Central railroad tracks.
Note: back cars in 1 at a time. Please cooperate. This stop involves a moderate hike to the top of Tom's Hill (a climb of 600 vertical feet).

Exposures to be encountered during this climb include the previously examined schistose marble and schists of the Walloomsac Formation. Rocks exposed at the higher elevations on this hill are tentatively assigned to the uppermost Walloomsac Formation. These schists are dark gray to silvery gray, contain staurolite, garnet, biotite, plagioclase, muscovite, and quartz, and locally have milky white quartz pods and stringers parallel to the schistosity.

The contact between this schist and the underlying Walloomsac units is interpreted to be conformable. However, the rocks resemble the Everett Formation as exposed at June Mountain, and detailed mapping is being undertaken to see if any evidence exists for a faulted contact.

The dominant fold structure here is an overturned synform that has been refolded (see Fig. 7).

Return to cars. Turn left onto Twin Lakes Road.

51.2 Intersection with Weatogue Road. Turn right.
52.0 Intersection with Rt. 44. Bear left.
52.3 Cross Housatonic River. Continue straight across intersection with Rts. 126 & 44
52.8 Turn left onto Sand Road. Unit B outcrops in field to left.
54.1 Intersection with Rt. 7. Turn left on 7. Outcrops of Walloomsac schistose marble along right.
54.7 Turn right onto Lower Road.
55.6 STOP 8. Pfizer Quarry. Unit A is quarried here. Depending upon quarry activities blocks with well-developed tremolite blades may be exposed. Refer to Figure 8 for the local structural relationships.

Turn around, proceed northwest on Lower Road.

56.5 Rt. 7. Turn right.
56.6 Continue straight across Rts. 7 and 44 intersection.
57.6 Merge with Rt. 7. Keep to right and travel north on 7.
58.4 Sheffield town line.
60.4 Cross Housatonic River.
63.2 Sheffield town center.
64.2 Turn left onto Cook Road.
64.4 Cross railroad tracks.
64.6 Cross Hubbard Brook. Keep straight.
64.8 Y-intersection with Bow Wow Road. Keep to left on Cook Road.
65.1 STOP 9. Bears Den. Park on right side of road. Cross private property. This exposure reveals an apparently conformable contact between the Walloomsac Formation and Unit C of the Stockbridge Formation. Within the Ashley Falls quadrangle the Walloomsac rests on all units of the Stockbridge Formation indicating a major unconformity beneath the Walloomsac.

Unit C is a fine-grained, calcitic dolomite marble that weathers light gray or dark, steel-gray. It may contain thin, phlogopitic partings and thin laminations of blue-gray dolomitic marble.

End of trip. Proceed to Rt. 7. Drive north on Rt. 7 to Monument Mountain Regional High School (driving time approximately 15 minutes).

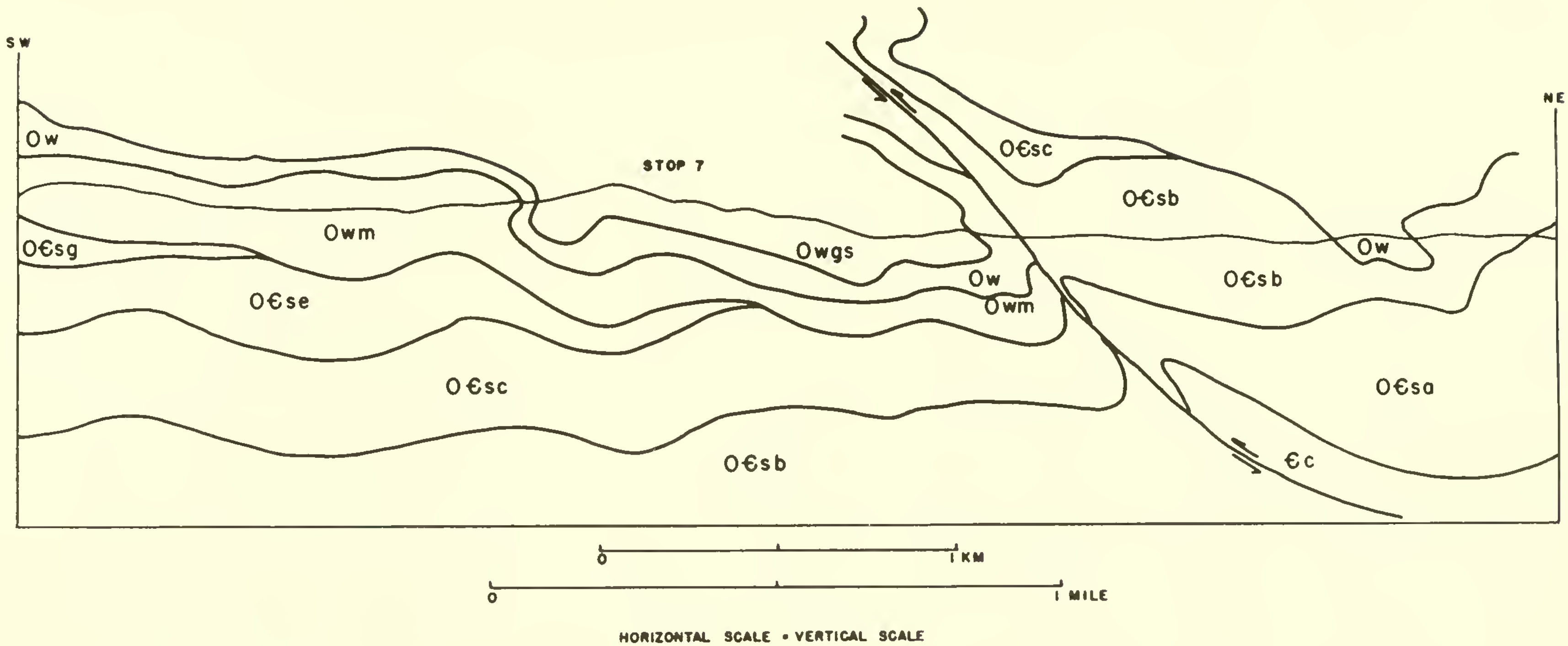

Figure 7. Generalized geologic section through area near Tom's Hill (Stop 7) (after Ratcliffe and Burger, 1975).

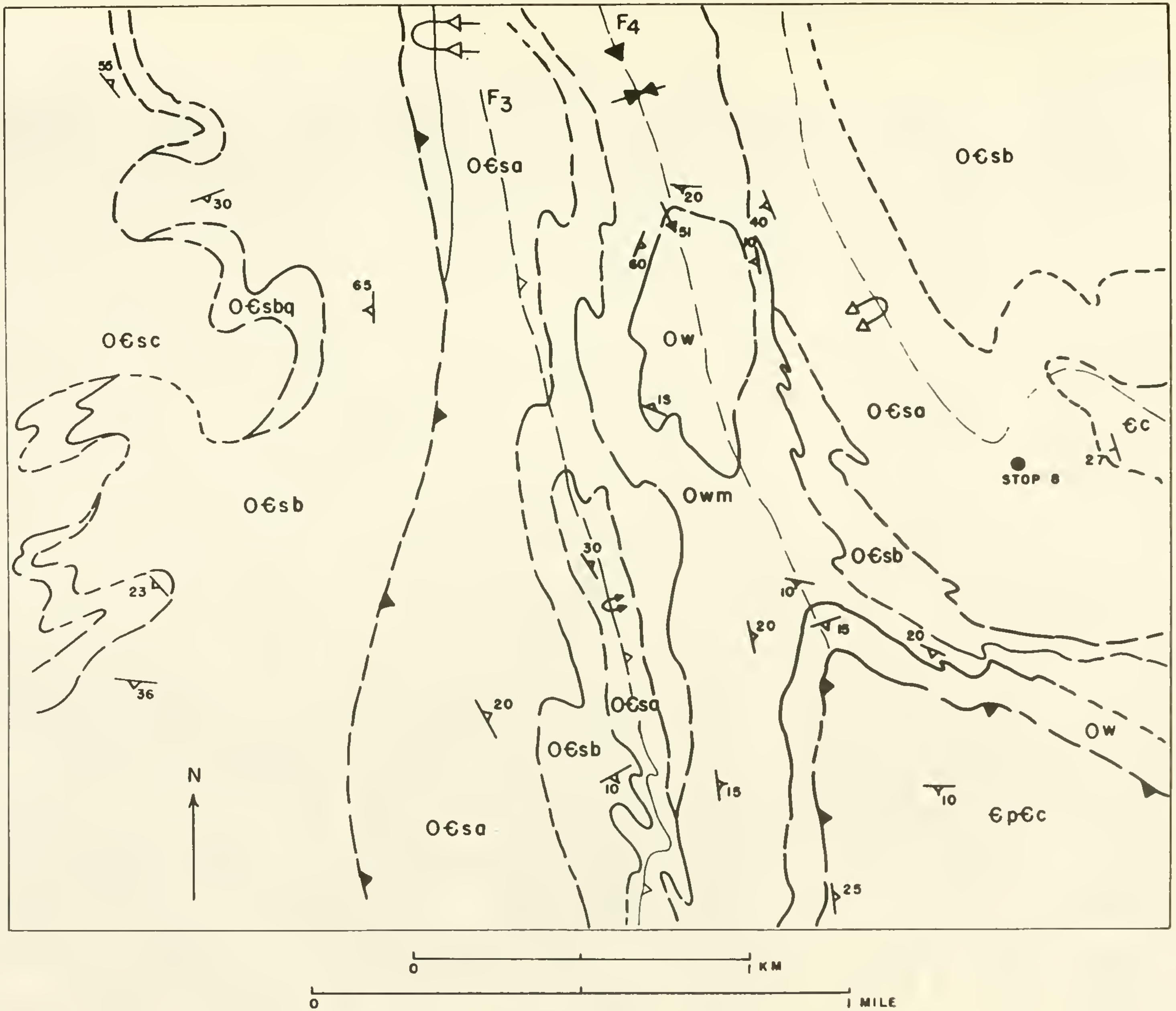

Figure 8. Generalized geologic map of the Canaan Mountain area (Stop 8) (after Ratcliffe and Burger, 1975).

BOULDER TRAINS IN WESTERN MASSACHUSETTS - REVISITED

George C. Kelley and Walter S. Newman

The New England landscape is strewn with numerous concentrations of rock fragments ranging from fine materials to blocks of house size. Close examination of these loose rocks in the past has sparked the imagination of investigators and elicited what today may be considered bizarre hypotheses concerning their origin. Specific interest in the nineteenth century was focused on the significance of erratic rock fragments in the interpretation of drift sheets. Writers have attributed the emplacement of erratics observed in New England to a variety of mechanisms, including pack ice, icebergs, waves of translation, streams, and even glaciers.

Rock fragments which differ in lithology from the underlying bedrock are, by definition, erratics, regardless of the mode of transportation or deposition. Erratics correlated with specific rock outcrops are indicators which typically occur as particle concentrations extending in linear or fan-shaped dispersions from a source area. During the nineteenth century, these features were collectively called "Boulder Trains," regardless of their geometry or particle size. Today they are more specifically called "indicator fans" or "indicator trains" (Flint, 1971).

Numerous boulder trains have been observed in New England (Flint, 1971), but few have the historical interest achieved by the Richmond Boulder Train in western Massachusetts. It was first described by Stephen Reed in an 1842 article in the local Lenox Farmer, and then in 1845 in a paper he read to the Association of American Geologists. He identified this train as "a chain of erratic, serpentine rocks." He noted the distinctively large size of the boulders, their tough, resistant lithology, a train length of approximately 20 miles trending to the southwest, an average breadth of 20 rods, and that the largest boulders were on the southeast-facing slopes and had crossed hills higher than their source area on Canaan Ridge.

The Richmond Boulder Train is important because its initial investigation and description coincided with the emergence of the glacial theory in North America, and the concomitant controversy regarding the deposition of drift sheets. The accessibility of this train, the exposure afforded by open farm fields, and the distinctive placement of boulders on the terrain surface, encouraged investigation by several eminent geologists, including Sir Charles Lyell, Edward Hitchcock, James Hall, and Louis Agassiz.

Hitchcock (1844; 1845) extended Reed's (1845) observations by noting that the train was longer than 29 meters, and that there were actually two trains. He was the first to publish speculations on the mode of deposition, although he did not propose any explanation. He rejected water currents because of the large boulders, the straight course of the train, and the lack of boulder rounding. He rejected icebergs because boulder emplacement would have required the successive movement of many large boulders along a narrow line. He rejected both glaciers and packed river ice because he doubted that either existed here. His rejection of glaciers seems inconsistent with several of his own concepts regarding drift deposition. In 1841 Hitchcock obviously envisioned continental glaciation. He postulated that ice perhaps 1200 meters thick covered the entire area. This ice was responsible for drift, for the transport and emplcement of large blocks and gravels over wide areas, and for the emplacement of boulders upon the crests of narrow mountains (1841, p. 251-256). By 1845, however, he seems to emphasize only alpine glaciation, comparing the blocks in the Richmond Boulder Train to Agassiz' alpine moraines, and stating:

> "But when we come to examine the country with reference to a glacier, we shall find it about as difficult to imagine the existence of one there as of a river. ...if we can imagine a glacier to start from the ridge in Canaan, it must ascend 100 or 200 feet...in order to go over the next ridge into Richmond;..." (Hitchcock, 1845, p. 264-265)

An additional explanation for these unusual boulder deposits was summarized by G. William Holmes (1966) in his salutation to Stephen Reed. This explanation involved a "wave of translation" transporting and depositing the boulders, and was first advocated by Rogers and Rogers in 1845. Holmes' summary states (1966, p. 434-435):

> "Rejecting glacial transport and iceberg drifting, they held that the train was the result of a sudden discharge of the Arctic Ocean southward, as a wave of translation (not unlike a tsunami). The Arctic waters picked up speed sweeping down the south-west slopes of the Adirondack Mountains and drove enormous ice islands against the summit of The Knob. This produced a vortex 'endowed with an excess gyratory or spiral velocity' which had a pendant column (similar to a tornado funnel or dust devil in the atmosphere). The whirlpool then gathered into its rotating column blocks from the summit and strewed them in a line along which its pendant apex dragged along the ground."

Desor (1848) was the first to formally propose a glacial origin, noting that, in Switzerland, similar trains existed which extended in the direction of glacial movement. This idea, however, was not accepted by the scientific community in America for more than two decades.

During those decades, Sir Charles Lyell entered the debate (1855; 1871). He traversed the Richmond Boulder Trains, and defined five additional linear trains on the landscape parallel to Reed's (1845) main boulder train. These additional trains consisted chiefly of carbonate rocks. Lyell attributed their source to the Richmond Range lying southeast of, and parallel to, the Canaan Ridge, which was the metavolcanic source for Reed's main boulder train. Since carbonate rocks are extensively exposed in this region, the carbonate trains (Lyell, 1855; 1871) are not "indicator trains."

Although Lyell (1871) was specifically investigating "glacial drift and erratics" in Berkshire County, Massachusetts and adjacent parts of New York, he rejected glacial emplacement for these blocks. He believed that glacially deposited blocks "would have radiated in all directions from a centre, whereas not one even of the smaller ones is found to the westward..." (p. 358). He conjectured "that the erratics were conveyed to the places they now occupy by coast-ice, when the country was submerged beneath the waters of a sea cooled by icebergs coming annually from arctic regions." (p. 361).

A glacial origin was again advocated in 1871, this time by John Perry. He visualized thinning ice, perhaps 200 meters thick, moving around nunataks such as Douglas Knob. This ice received loose blocks on the surface which were subsequently deposited downstream as the ice melted. A year later Louis Agassiz (1872) presented his views and effectively ended the debate. He also postulated glacial origin for the trains, but he believed a large ice sheet 3000-3700 meters thick was responsible. Reed (1873) presented further observations supporting glacial transport for the second boulder train.

The most detailed study of the trains was made by Benton in 1878. He confirmed the observations of Reed (1845; 1871), but found fragments of only two of Lyell's (1855; 1871) five carbonate trains.

About 30 years later, Frank Taylor (1910) reconnoitered this region. He found intermittent "amphibolite" boulders extending southward to State Line, Alford, and Great Barrington, which he named the Great Barrington Boulder Train. He believed these scattered fragments represented an early ice invasion which incorporated talus material into the basal part of southward-flowing ice. This mode of origin and transport would account for the greater degree of weathering and rounding of the boulders than observed in the Richmond trains, and also would account for their partial burial in till. Recent mapping (Ratcliffe, 1974 personal communication) revealed other meta-volcanic rock exposures in ridges south and east of Canaan Mountain. Other lithologies occurring to the west of this area may be mistaken for the meta-volcanic rocks exposed on Canaan Mountain. The reconnaissance level at which Taylor (1910) made his observations, the possible lithologic confusion, the additional metavolcanic source areas, and the notable gaps along the route of the train, raise doubts about the existence and significance of Taylor's (1910) "Great Barrington Boulder Train," and its relationship to the Richmond Boulder Trains.

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ROAD LOG

TOPOGRAPHIC QUADRANGLES (1:24,000, 7-1/2 minute) - Great Barrington, Egremont, State Line, Canaan, Pittsfield West, and Stockbridge

0.0 Start at Monument Mountain School, proceed south on Route 7 to Great Barrington, and through the main business district.

4.0 Turn right on Alford Road (Castle Street). Proceed northwest, driving through a wooded, suburban area which has thin till with numerous bedrock exposures.

5.2 The field to the left of the road contains many unweathered boulders of Cheshire quartzite. Boulders of this same lithology and a similar population density occur in and beyond the orchards on the right (north).

5.7 An outwash apron originating from a head-of-outwash ice-contact position a short distance to the north has apparently covered the quartzite boulders in the vicinity of the Simon's Rock School.

6.3 Intersection, Seekonik Road. Quartzite boulders are again evident to the right, while those presumably present to the left of the road are covered by Holocene alluvium.

6.8 Deusenville Road. High density of quartzite boulders to the right.

7.2 STOP 1 - CHESHIRE QUARTZITE EXPOSURE

(Park along the right shoulder. Please EXERCISE CAUTION when exiting and entering your vehicle here and at all stops)

The Cheshire quartzite crops out for about a mile along a north-striking ridge and east-facing escarpment. The quartzite is well exposed, starting about 100 feet above the level of the road. Microstriations strike within 15 degrees of N. 60° W. (use lead pencils to reveal those scratches!). These exposures appear to be the source of the quartzite erratics found to the east and southeast of this stop. However, since the Cheshire quartzite also crops out 1.4 miles to the northeast, and 1.6 miles to the southeast, this area is by no means a unique source of quartzite erratics. Nevertheless, it is presumed that the boulders we drove by between Great Barrington and here had their source in these exposures, since the lithology of the boulders is identical to that of the supposed source. Cheshire quartzite boulders are not evident west of this area. The strong easterly component of glacial flow demonstrated here (and as will be demonstrated again later today), across the topographic grain of the landscape, suggests glacial outflow controlled by the axis of the Hudson-Champlain Lowland (Kelley, 1975, p. 234-242 of this volume) or a more remote center of outflow far to the northwest.

Continue northwest on Alford Road.

8.6 At intersection, bear left into Alford.

8.8 Bear right onto West Street. Area is bedrock covered by thin till.

9.8 Enter State Line quadrangle.

12.6 Intersection, turn left on West Center Road.

14.6 Extensive dead-ice and ice-contact topography.

15.5 Intersection, turn left on Route 102. A well-defined kame terrace abuts the north slope of this pass through the Taconic Range.

16.5 Turn right onto Route 22. The gradient of the kame terrace is still apparent on the right and suggests water flow was to the east.
18.5 Enter Canaan quadrangle.
19.6 Large erratic may be seen to the left of the highway.
21.9 Queechy Lake to the left. Observe ice-contact topography evident along this valley in the next three miles.
22.6 Berkshire School. Reed's second (southwestern) boulder train extends across this valley, through the school campus, from northwest to southeast. Scattered boulders are evident on both sides of the road.
23.2 Reed's main boulder train crosses Route 22 from the northwest to southeast. (The trip will return to this area later)
23:3 An ice-channel filling is exposed in the gravel pit to the right.
24.6 STOP 2 - SMALL MORAINE

(Route 22 is often busy, and cars move rapidly. The caravan needs to turn around in driveways north of this moraine, at points where visibility allows safe turning. Return southward after turning, park on the west shoulder. USE CAUTION EXITING/ENTERING CARS)

The formation of the landscape features and deposition of materials associated with this moraine will be considered. Problems include the origin of abandoned drainage channels, and the gradient on outwash materials south of this moraine.

Continuing south along Route 22, Douglas Knob becomes visible to the southwest through the trees. It is the highest peak on the Canaan Mountain ridge and is considered the point source for the metavolcanic erratics comprising Reed's main and second boulder trains.

26.0 STOP 3 - REED'S MAIN BOULDER TRAIN

(Park on west side, and cross the small South Wyomanock stream)

Significant aspects of the boulders and their placement on the terrain can be examined. Points to ponder - What is the degree of boulder surface weathering? Are the boulders rounded or angular? How "tough" is the boulder material? What is the average size of the fragments? Is there any apparent orientation among these fragments? Is it significant? How well defined are the lateral limits for this train? Why? Was Douglas Knob the point source for both of Reed's trains (there is a gap in the boulders of the second train in the vicinity of the swamp north of Queechy Lake)?

26.6 STOP 4 - REED'S SECOND BOULDER TRAIN

Boulders in this train can be compared with boulders in Reed's main train. Factors of weathering, rounding, burial, orientation, size and continuity should be considered. Were these trains emplaced simultaneously? When in the glacial cycle did emplacement occur?

Proceed south on Route 22.

28.0 Turn left (southeast) onto Route 295. Note the paucity of boulders on the observable terrain.
30.1 Enter Pittsfield West quadrangle.
30.2 Reed's second train crosses Route 295 between this point and the next intersection at Richmond Road. Where are the boulders? The second train is often poorly defined, and distinct gaps exist between concentrations of erratics.
30.5 Turn left (north) onto Richmond Road.
31.1 STOP 5 - LYELL'S ROCK

Reed's main boulder train crosses Richmond Road and, unlike the paucity of boulders in the secondary train, numerous boulders are evident on each side of the highway. This location is on a lee slope, where boulders

tend to have a larger mean size than do boulders on the stoss slopes. Cross the highway to the northwest, enter the woods, contemplate the orientation of the particles, seek the boundaries of the train, and then proceed on foot "up-ice" to a large "particle" (16 x 6 x 6 meters) which is commonly called Lyell's Rock. Lyell's sketch of this rock indicates a broken fragment derived from the nearby, larger block. Other boulders were once joined to adjacent or nearby fragments. Perhaps initial size, transport, emplacement, and post-depositional shattering account for the downstream decrease in mean particle size. Lyell's Rock is the second largest erratic in these trains. The largest, called "The Alderman," is 29 meters long. It is on Dupey's Mount in the Richmond Range of the Taconic Mountains (Canaan quadrangle).

Proceed northeast on Richmond Road.

31.7 Turn right (south) onto Dublin Road. Observe the frequent carbonate erratics in the adjacent field. These are believed to be part of Benton's minor carbonate train, lying northeast of the main train.
32.5 Turn right (west) onto Summit Road.
32.6 Reed's main boulder train crosses Summit Road.
32.9 Carbonate erratics evident to the right of the road
33.2 Turn left (south) on Route 41.
33.5 Reed's second boulder train crosses Route 41.
33.7 Turn left (west) onto Sleepy Hollow Road. Reed's second train extends towards the southeast across this road.
34.0 Carbonate rocks are again evident.
34.3 Re-cross the path of Reed's main boulder train.
34.4 Cross railroad tracks. Observe metavolcanics in the field walls.
34.6 Turn right (south) onto Medlyn Road.
35.4 Oblique intersection of Swamp Road. Continue south on Swamp Road.
35.8 STOP 6 - RECAPITULATION
Reed's main train extends across this terrain toward the southwest, where it crosses Lenox Mountain. Mean size of boulders is distinctly smaller in this region than upstream. Farmers have removed the boulders from fields to adjacent walls, a procedure which becomes more significant to the appearance of the train farther downstream. Reed's second boulder train is 0.2 miles south of this location.

This is the last trip stop. Those returning to the Monument Mountain School should continue southward on Swamp Road into the village of West Stockbridge.

39.0 Intersection with Route 102. Continue straight southward on Route 102 to Stockbridge
44.0 Turn right (south) on Route 7.
46.6 Enter Great Barrington quadrangle.
47.6 Monument Mountain School

Stratigraphy and Structural Geology in the Amenia-Pawling Valley, Dutchess County, New York

by

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Introduction

The purpose of this field trip is to provide familiarity with the stratigraphy and structure of the Amenia-Pawling portion of the Harlem Valley in eastern New York State. Emphasis is placed upon stratigraphic relations of the Wappinger Group (Dana, 1879) which represent the Cambrian-Ordovician carbonate shelf sequence in this portion of the Appalachians. Additional stops will be made to examine the Poughquag Quartzite, the Walloomsac Schist, the Everett Schist, and Precambrian units of the Housatonic Highlands, all of which are integrally related to the regional geology.

It should be understood that the structural framework and geologic history of this area have not been clearly deciphered. As in other parts of the Taconide Zone (Zen, 1972), complex polyphase deformation, regional metamorphism, and limited exposure have combined to leave scanty evidence of a protracted geologic history. A great deal more detailed work is required before the Harlem Valley area is well understood. It is our hope that this trip may arouse sufficient interest in the regional geology so that others will decide to undertake further field studies in the area.

Acknowledgements

We thank the following individuals for their roles in enhancing our understanding and interest in the local geology: John Rodgers, Yngvar Isachsen, and Rosemary Vidale. One of us (JMM) was supported by a NSF Science Faculty Fellowship during the summer of 1971.

Previous Work

Early work concerning the carbonate stratigraphy was conducted by Dana (1879), Dwight (1887), Mather (1843), Merrill (1890), Walcott (1891), and Dale (1923). Dana (1879) named the Cambrian-Ordovician carbonate units the Wappinger Group. Although the term Stockbridge Formation (Emmons, 1842) has precedence, Dana's terminology has generally been applied within New York State. We shall adhere to this tradition.

Dale (1923) mapped the carbonate rocks of western Connecticut and eastern New York and showed that units of the Harlem Valley could be carried through to the Stockbridge Valley of western Massachusetts. He divided the carbonates into lower dolomitic and upper calcitic sequences.

The most important contribution to the stratigraphy of the carbonate rocks was carried out by Knopf (1927, 1946, 1962) in the terrain around Stissing Mt., New York. Her subdivision of the Wappinger Group is the one adopted in this study and is presented in correlation chart form in table 1.

Balk (1936) carried out an extensive study of the structural geology of Dutchess County. He considered subdividing the carbonates but ultimately decided not to do so on the ground that intense deformation and metamorphism made the attempt impractical.

Carroll (1953), working in the Dover Plains 7½' quadrangle recognized an upper (western) calcdolomite and dolomite section and a lower (eastern) dolomite section.

Waldbaum (1960) mapped the valley carbonates between Dover Plains and Wingdale, N.Y., and his subdivisions correspond approximately to Knopf's (1927, 1946, 1962). For the most part, his contacts approximate our own (fig. 2).

The pelitic rocks of the area were investigated by Balk (1936) who classified them as Hudson River pelites of Cambrian-Ordovician age. He was unable to subdivide these units, and, to a great extent, this stratigraphic uncertainty remains today. As a result, tentative correlations are made to less metamorphosed, or better understood units, outside of the area--i.e. Walloomsac Slate (Prindle and Knopf, 1932); Everett Schist (Hobbs, 1893); Manhattan A, B, C (Hall, 1968).

The structural geology of the area was considered in detail by Balk (1936). He concluded, largely on the basis of minor structures, that none of the rocks in the area were allochthonous, but that numerous reverse faults brought older rocks up against younger ones. Both Carroll (1953) and Waldbaum (1960) reached similar conclusions, but held open the possibility of far-traveled thrust slices. Carroll (1953) considered the presence of retrograde metamorphism in some of the metapelites to be suggestive of an allochthonus history.

In preparing the 1961 and 1973 editions of the New York State Geological Map, Fisher demonstrated the existence of Taconic "soft-rock" allochtons (gravity slides) to the west in the general vicinity of the Poughkeepsie, N.Y. "Hard rock" slices of Everett Schist were recognized in the Dover Plains and Millbrook 15' quadrangles (Fisher et al., 1973). It thus appears that allochthonous rocks are widely represented in the surrounding area and extend into the region considered in this report (fig. 2). Parautochthonous carbonate rocks and gneisses are also recognized by Fisher and Warthin (unpublished) in western Dutchess County. D. W. Fisher and A. S. Warthin, Jr. have, in preparation, a text and colored geologic maps of the western half of Dutchess County, New York.

Metamorphism

The Amenia-Pawling Valley represents the eastern section of a classic sequence of progressive Barrovian metamorphism (Balk, 1936; Barth, 1936).

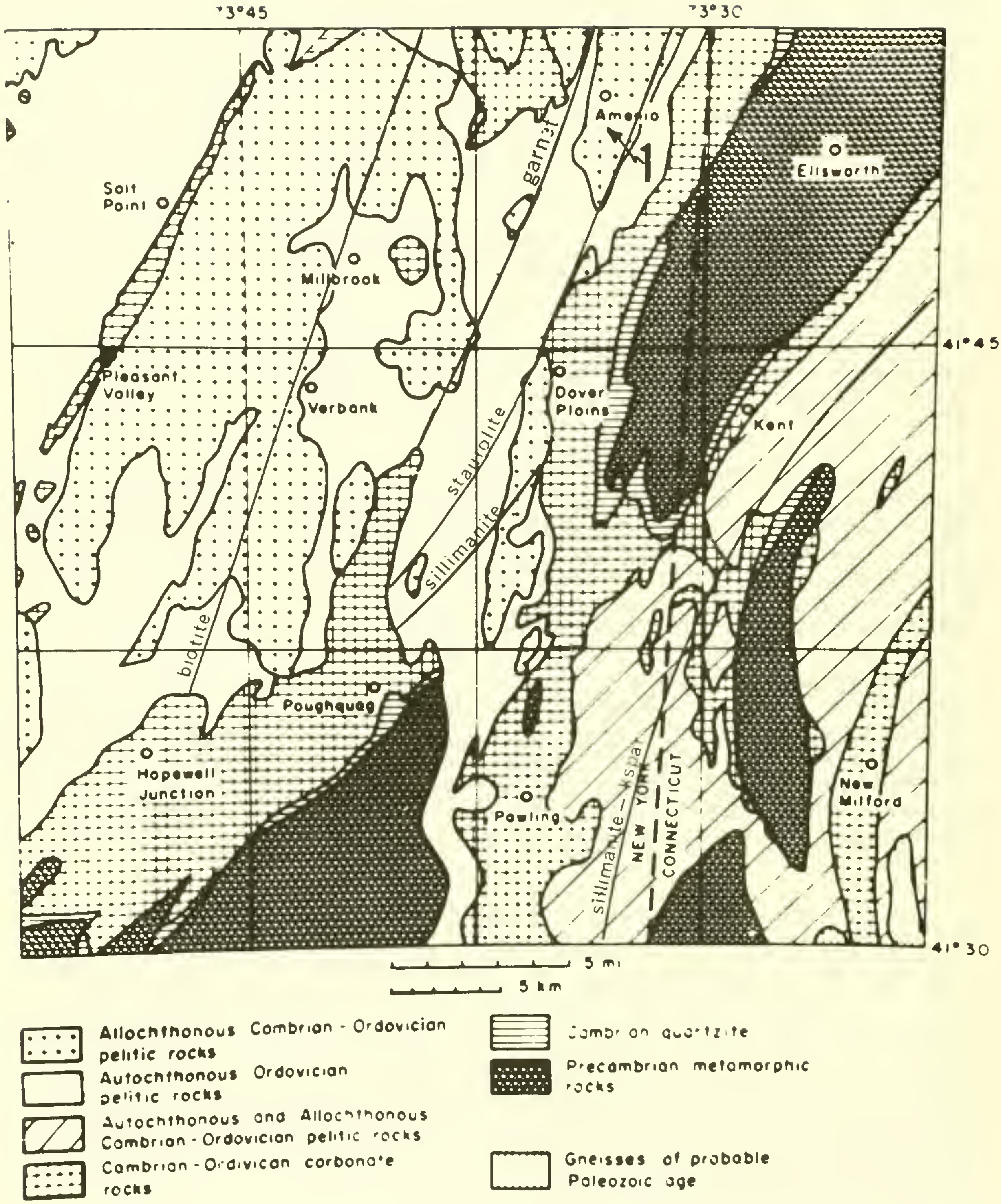


Fig. 1 - Generalized geology of the Amenia-Pawling Valley, positioning of isograds, and location of Stop 1. (After Vidale, Geol. Soc. America Bull., v. 85, 1974).

In recent years the isograds and metamorphism have been studied by Rosemary Vidale (1974). The progressive nature of the metamorphism is not well displayed in the Harlem Valley, because its trend is approximately parallel to the metamorphic isograds (fig. 1).

Although the petrological aspects of the metamorphism have received considerable attention, uncertainty continues to exist concerning its age. To the north, in areas mapped by Zen (1969) and Ratcliffe (1969), the higher grade isograds have been assigned a Devonian age (Acadian Orogeny). To the south, Long (1962) and Ratcliffe (1967) demonstrated that the 435 m.y. old Cortland Complex transects the metamorphic rocks and that the metamorphic events may be associated with a Taconian (∿450 mya) metamorphism. This is consistent with ∿400 mya Rb/Sr ages in the Walloomsac near Verplank, N.Y. (Long, 1962). A set of younger Rb/Sr ages clustering around 350 my indicates an Acadian overprinting of the Taconian metamorphism. According to Long (1962), this overprinting increases eastward in its intensity. The development of two generations of biotite reflects the overprinting. In considering similar problems in the Manhattan Prong area; Hall (1968) left open the possibility of either a Late Ordovician (Taconian Orogeny) or Middle Devonian (Acadian Orogeny) age for the peak metamorphism. Within Dutchess County similar uncertainty exists.

Rock Units and Stratigraphic Detail

(I) Precambrian Basement

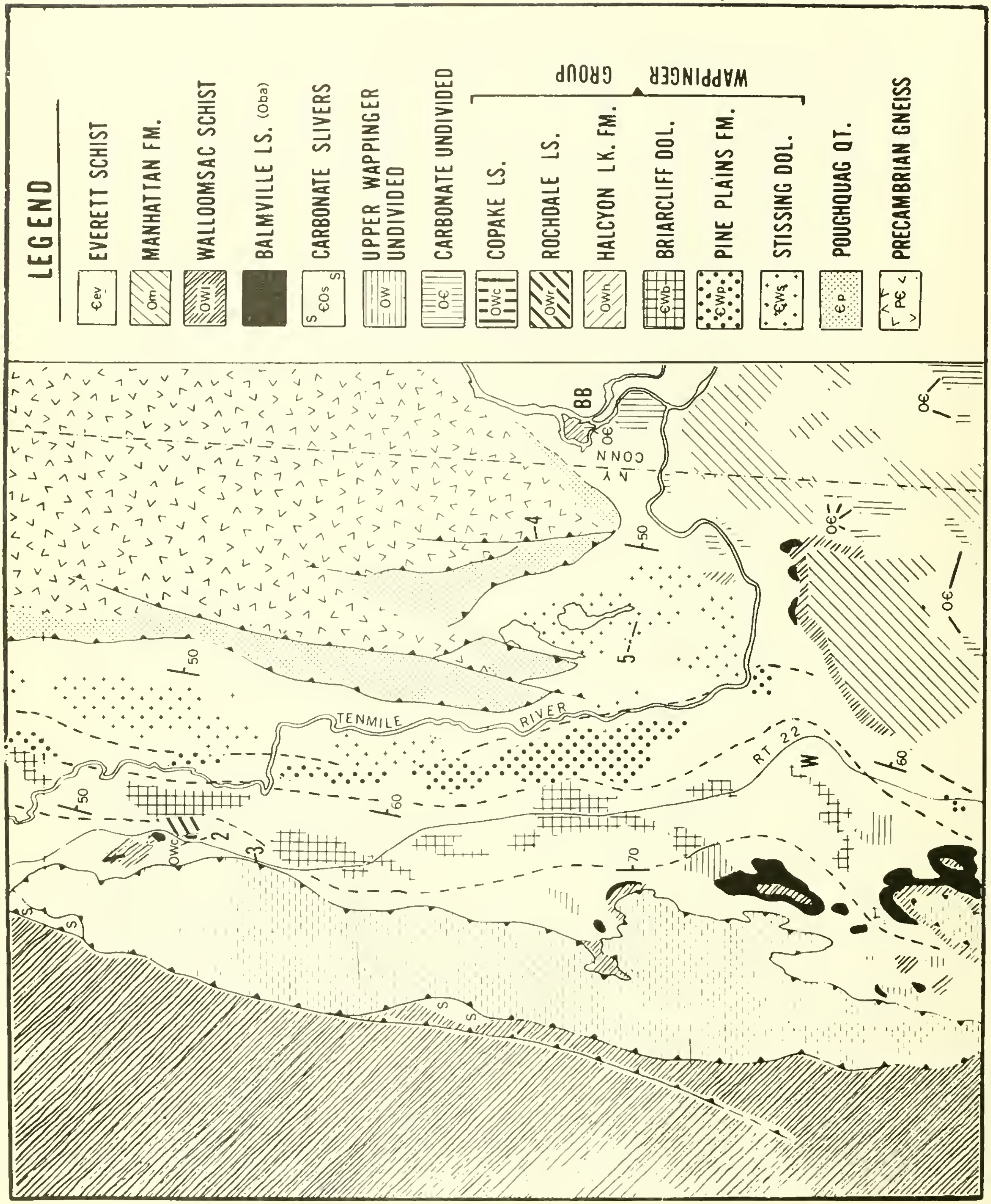
Rocks of the Proterozoic (Helikian) basement are exposed in three areas (1) Corbin Hill, (2) Housatonic Highlands, and (3) Hudson Highlands. Quartzofeldspathic gneisses, biotite-quartz-feldspar gneisses, and amphibolites dominate these units, although other lithologies are present.

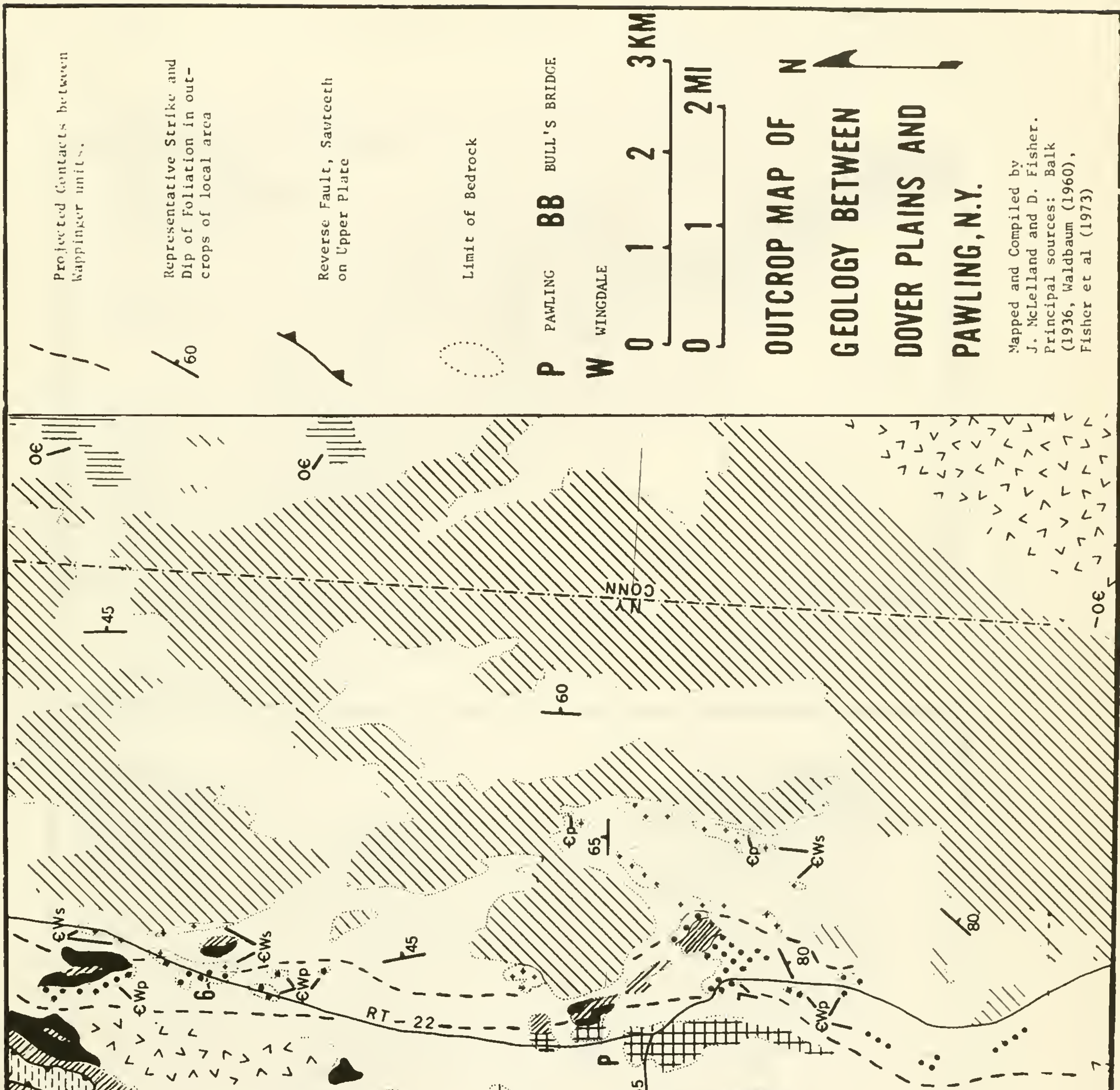
(II) Poughquag Quartzite (Dana, 1872) (∿50-200 m)

White, tan, and pink, massively bedded vitreous quartzite. Throughout the area it appears to be relatively clean, but lower conglomeratic horizons have been recognized. Quartz content generally exceeds 90%. Bedding is rarely visible. Near contacts with the overlying Stissing carbonates rosettes of tremolite are developed. Rapid gradation into the Stissing is achieved by interlayering of quartzite and quartz bearing dolostones over a stratigraphic distance of 10-15 meters. In places the Poughquag lies unconformably upon Proterozoic basement gneisses--i.e. East Mt. (Waldbaum, 1960). Often the contact is marked by reverse or high angle normal faults. Early Cambrian olenellid trilobites have been identified in western Dutchess County, N.Y.

(III) Wappinger Group (Dana, 1879) (∿1000-1500 m)

The Wappinger Group comprises the Cambrian-Early Ordovician carbonate shelf sequence in the Hudson Valley of New York State. It is equivalent to the Stockbridge Formation (Emmons, 1842) in Massachusetts and Connecticut (Table 1). The following subdivisions can be recognized. They are listed in order of decreasing age.


LEGEND
EVERETT SCHIST
MANHATTAN FM.
WALLOOMSAC SCHIST
BALMVILLE LS. (Oba)
CARBONATE SLIVERS
UPPER WAPPINGER UNDIVIDED
CARBONATE UNDIVIDED
COPAKE LS.
ROCHDALE LS.
HALCYON LK. FM.
BRIARCLIFF DOL.
PINE PLAINS FM.
STISSING DOL.
POUGHQUAG QT.
PRECAMBRIAN GNEISS
WAPPINGER GROUP
TENMILE RIVER
RT 22
CONN
N Y
BB
W


Projected Contacts between Wappinger units.
Representative Strike and Dip of Foliation in outcrops of local area
Reverse Fault, Sawteeth on Upper Plate
Limit of Bedrock
P PAWLING
BB BULL'S BRIDGE
W WINGDALE
0 1 2 3KM
0 1 2MI
N
OUTCROP MAP OF GEOLOGY BETWEEN DOVER PLAINS AND PAWLING, N.Y.
Mapped and Compiled by J. McLelland and D. Fisher. Principal sources: Balk (1936, Waldbaum (1960), Fisher et al (1973)
N.Y.
CONN
RT-22
RT 55
P
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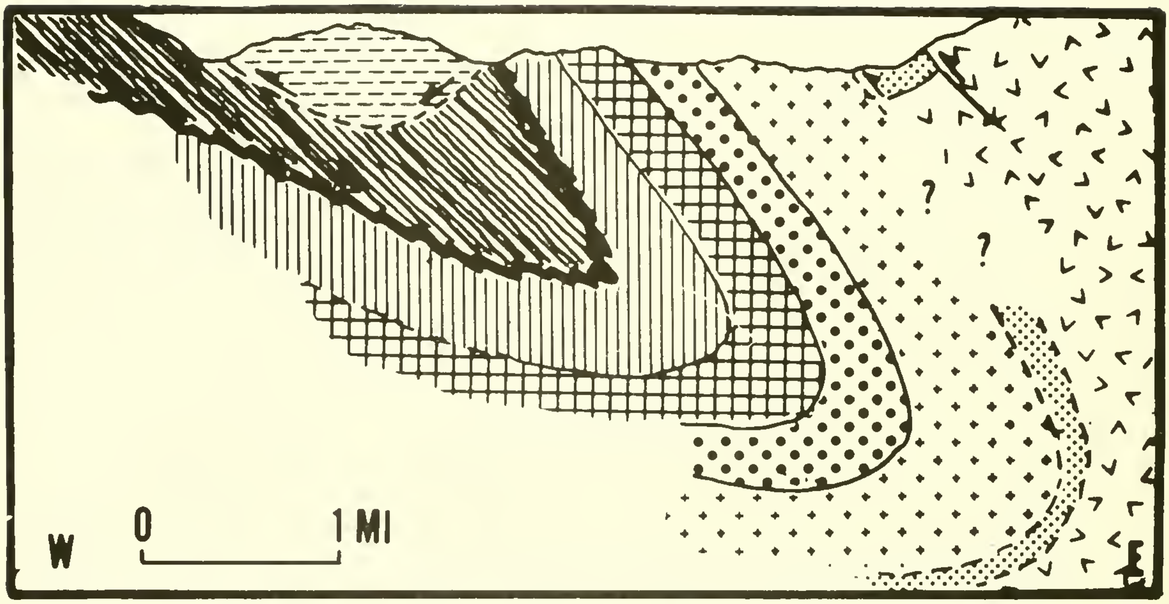

Fig. 3a - Schematic E-W cross section of Harlem Valley Syncline along a line through Nellie Hill. Symbols as in fig. 2.

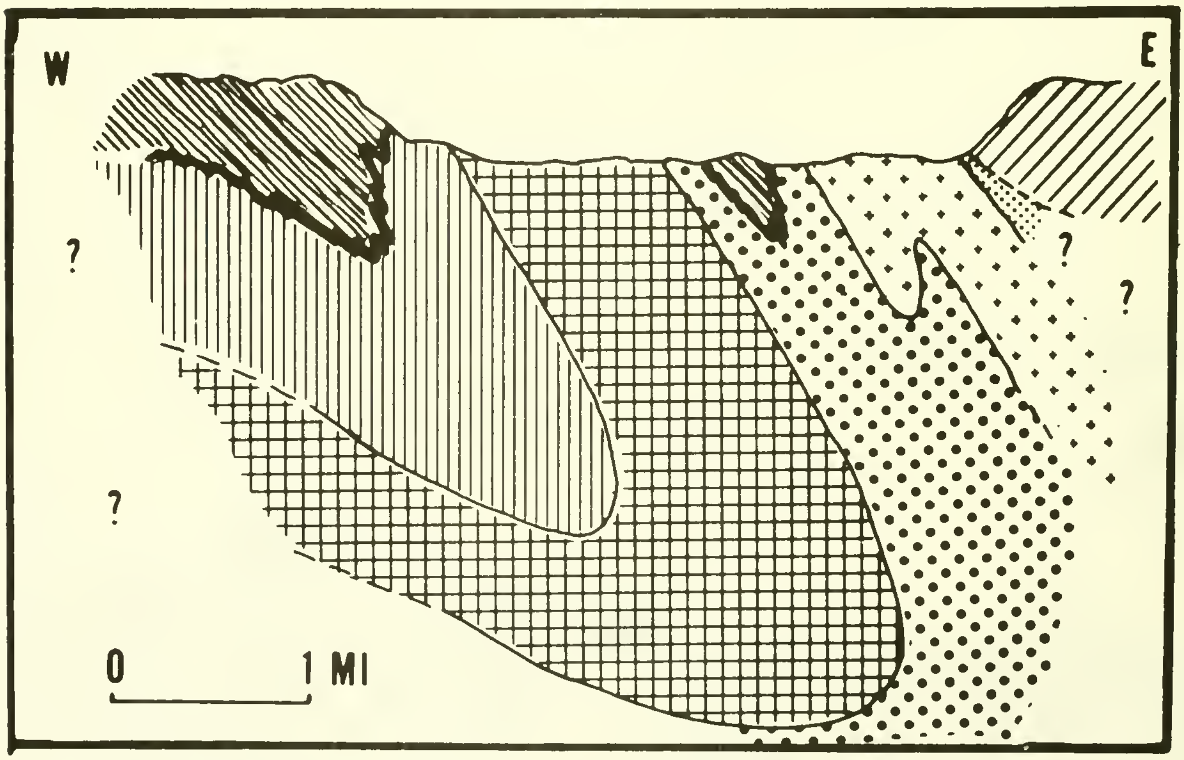

Fig. 3b - Schematic E-W cross section of Harlem Valley Syncline along a line subparallel to NY 55. Symbols as in fig. 2.

Age	Amenia-Pawling Valley		Bashbish Falls quadrangle State Line quadrangle (Zen and Hartshorn, 1966) (Ratcliffe, 1969)	
Earliest Cambrian or Proterozoic (Hadrynian)	Manhattan B, C, Schist Everett Schist		Everett Formation Nassau Formation	
Middle Ordovician	Walloomsac Schist Balmville Ls.		Walloomsac Formation Egremont Phyllite Limestone / Schistose Marble	
Early Ordovician	Wappinger Group	Copake Limestone	Stockbridge Formation	Unit g
		Rochdale Limestone		Unit f
		Halcyon Lake Fm.		Unit e
Late and Middle Cambrian		Briarcliff Dolostone		Unit d
				Unit c
		Pine Plains Fm.		Unit b
Early Cambrian		Stissing Dolostone		Unit a
	Poughquag Quartzite		Cheshire Quartzite	
Cambrian ?			Dalton Formation	
Proterozoic (Helikian)	Gneisses at Corbin Hill and Housatonic Highlands			

Table 1. Chart Showing Correlation of Map Units - southeastern New York, southwestern Massachusetts, and northwestern Connecticut.

(A) Lower Wappinger Dolostone Sequence

(a) Stissing Dolostone (Walcott, 1891) (∿300 m)

Typically massive sparkling white dolostones and calcitic dolostones that show a limited reaction with dilute HCl; weathers a pale grey and readily decomposes into white dolomitic sands. Local horizons are rich in yellow to white bands of chert and quartzite which are usually boudinaged. Within the lower 10-20 meters increasing quantities of quartzite layers mark the transition into Poughquag Quartzite. Tremolite and diopside develop near the chert and quartzite beds. Pelitic intervals occur and Mrs. Knopf (1946) recognized a 20 m layer of red shale in the vicinity of Stissing Mt. Fossils in western Dutchess County denote an Early Cambrian age; the uppermost strata may be of Middle Cambrian age.

(b) Pine Plains Formation (Knopf, 1946) (∿300 m)

The Pine Plains Formation is characterized by its extreme variability. It is predominantly composed of dolostone but dark grey phyllite layers are common and lavender to purplish mica-rich mottlings are widespread. Layers of dolostone, dolomitic siltstone, and dolomitic sandstone alternate providing a distinctive array of extremely well bedded strata. The thickness of individual beds is variable, ranging between 3 m to 0.5 m. The most characteristic color of the weathered surface is buff to brown or tan. Gray weathering, relatively pure, siliceous dolomites also appear in the section and are marked by the development of diopside and tremolite. Chert and quartzite beds are common in the well bedded portions and give rise to excellent examples of boundinage. At lower metamorphic grade, oolites, cross-bedding, ripple marks, and dessication cracks can be found in the Pine Plains Fm. Local graded bedding is associated with quartz grains in the dolostone.

(c) Briarcliff Dolostone (Knopf, 1946) (∿300-400 m)

A gray weathering, massive, light gray to dark gray dolostone containing abundant, and boudinaged, yellow to white chert bands. Minor pelitic mottling is present in some layers. Within this area the Briarcliff is relatively free of quartz sand and calcite. Weathers to rounded pavement outcrops in the field. Weathered-out knots of quartz are common and diagnostic. Diopside tablets and tremolite rosettes show abundant development parallel to siliceous layers. Disharmonic folding occurs between dolostone and cherty layers. Rare fossils near Pine Plains indicate a Late Cambrian (Trempealeau) age. The Briarcliff Dolostone is the thickest unit within the Wappinger Group.

(B) Upper Wappinger Sequence of Calcitic Marbles (∿300-400 m)

Poor outcrop and unconformable overlap by the Walloomsac/Balmville lithologies, have made it difficult to subdivide these units in the field. As a result, they are mostly mapped together as a single Upper Wappinger unit in fig. 2. Fortunately, stop 2 at Nellie Hill provides an excellent cut through portions of the Copake and Rochdale Limestones.

(a) Halcyon Lake Formation (Knopf, 1946)

Fine to medium grained calcitic dolostone with some chert. The base is usually sandy or silty. The Halcyon Lake Formation has proven exceedingly difficult to find and map in this area, and we have not yet recognized any lithology that can be definitely assigned to the Halcyon Lake Formation. Elsewhere in Dutchess and Orange Counties, fossils indicate an Early Ordovician (Gasconadian) age.

(b) Rochdale Limestone (Dwight, 1887)

The lower portion consists of interbedded buff-weathering, fine textured dolostones and calcitic dolostones. The upper portion contains purer, only slightly dolomitic limestones; some of these possess coarse textures. Buff to fair weathering sandy-beds are common, frequently displaying sedimentary textures. In western Dutchess County, fossils denote an Early Ordovician (Roubidoux an) age.

(c) Copake Limestone (Dana, 1879)

Grey to white weathering dolomitic limestone, coarse textured limestone, and dolostone. Basal portion contains sand and silt that tends to occur in pods and lenses giving the rock a mottled appearance. Cross-bedding is frequently developed in the sandy layers. Rare fossils elsewhere in Dutchess County are of Early Ordovician (Cassinian) age.

(IV) Balmville Limestone (Holzwaswer, 1926) (0-30 m)

The Balmville consists of a coarse textured, blue-gray weathering calcite marble that is free of dolostone layers. Conglomeratic clasts of underlying Wappinger carbonates are relatively common. Locally the marble is schistose. It grades upward by interdigitation into the black Walloomsac phyllites. Layers of calcite bearing calc-silicate-biotite-quartz-plagioclase rocks are commonly developed in the transition zone. The Balmville Limestone is not everywhere present at the base of the Walloomsac Schist, and this absence is probably due to local non-deposition. Elsewhere, the Balmville Limestone has been found resting upon different Wappinger units. Balmville fossils indicate correlation with Middle Ordovician Mohawkian (Rockland) units farther to the west.

(V) Walloomsac Schist (Prindle and Knopf, 1932) (∿500 m ?)

The original name of Walloomsac Slate was given to certain black slates overlying the Trenton-equivalent limestones in Rensselear County, N.Y. In Columbia and Dutchess Counties the term Walloomsac is applied to phyllite and schist equivalents of the Snake Hill shales farther to the west. Zen (1969) included the Balmville Limestone as the basal member of the phyllite/schist sequence and referred to the entire mass as the Walloomsac Formation.

The Walloomsac schists are typically jet black to rusty weathering phyllites that contain graphite and pyrite. Biotite tends to dominate over muscovite in the mode (see below). Many sections contain abundant quartz and plagioclase and tend to be more granulitic than schistose in texture.

Everett Schist (Hobbs, 1893) (∿700 m ?)

This unit was named for exposures on Everett Mt. in southwestern Massachusetts near the New York line. Typically the Everett consists of green-gray and silvery schists, phyllites, and green-tan massive quartzites. It tends to be quartz rich and to show abundant development of muscovite, garnet, and staurolite at high grade. Coarse muscovite dominates over biotite in almost all sections assigned with certainty to the Everett.

Within the Amenia-Pawling Valley the Everett Schist is the only allochthonous Paleozoic unit that has been recognized. Presumably it represents a hard-rock slice of later Taconian thrusting (Hudson Valley Phase of Taconian Orogeny). It is believed to be correlative with the Elizaville Argillite and Nassau Formation (Late Hadrynian or Early Cambrian) further west in Dutchess County.

Manhattan Schist (Merrill, 1890; Hall, 1968) (thickness indeterminable)

We have used this designation for rocks that cannot be placed with certainty within either the autochthonous Walloomsac Schist or the allochthonous slices of Everett Schist. These lithologies, which are exposed in the highlands to the east of Pawling and Wingdale, consist of micaceous schists containing abundant stringers and veins of quartz and quartzo-feldspathic material.

In its type area the Manhattan is divisible in a Lower (A) and Upper (C) unit separated by an amphibolite unit (B) (Hall, 1968). The lower unit consists of a dark, biotite rich, graphitic member. The presence of basal carbonate rich rocks strongly suggests that the lower Manhattan correlates with the Walloomsac Schist. The upper Manhattan consists of coarse, light colored muscovite schists. Garnet and staurolite are common. Ratcliffe and Knowles (1969) conducted modal analyses on 46 samples of Manhattan Schist. They report that out of 22 samples of Upper Manhattan (C), 19 show muscovite more plentiful than biotite; of 24 samples of Lower Manhattan (A), 19 show an excess of biotite over muscovite. Staurolite is present in 13 samples of upper Manhattan and is present (as small amounts) in only 6 samples of lower Manhattan. Opaques are much more abundant in the lower Manhattan than in the upper Manhattan unit.

It is believed, but unproven, that within the local area the upper Manhattan may be correlated with the allochthonous Everett Schist and the lower Manhattan with the autochthonous Walloomsac Schist. Similar suggestions were made by Hall (1968) for these units in the Manhattan Prong.

Structural Geology and Geologic History

(A) Chronology

As with other examples of Taconic geology, the region exhibits at least two, and frequently three significant deformational events of Paleozoic age. Following the terminology and scheme of Ratcliffe (1969), we have:

Deformational Event	Foliation	Tectonic Style
D_0-pre-Walloomsac Mid-Ordovician unconformity bevels down through, at least, the Stissing	None Recognized	Unknown. Possibly high high angle faulting, folding or both. Possible overturning of shelf sequence
D_1-Post Mid-Ordovician Unconformity Locally recognizable as refolded isoclinal minor folds cut by D_2 foliation. Vermontian Phase of Taconian Orogeny	None Recognized	Isoclinal recumbent minor folds. Related to emplacement of early allochthons (gravity slides)
D_2-Post and Pene-allochthonous Hudson Valley Phase of Taconian Orogeny	Major NE foliation (S_2)	Large recumbent folds that dominate structural framework, includes and post-dates hard-rock thrusting.
D_3-Post-S_2 Foliation Folding of S_2 Foliation. Probably Acadian in age.	Crenulation cleavage (S_3) and associated chevron folds. Trend varies from N-S to NNW.	Chevron folds, kink banding, microlithons along slip cleavage.

A fourth folding event is suggested by changes in plunge of lineation from north to south in several areas. They may be seen on Balk's 1936 geologic map of the Clove 15' quadrangle. Waldbaum (1960) mapped an E-W trending fold axis on this basis just south of Nellie Hill. These changes in plunge may reflect synchronous E-W cross-folding associated with the rise of the Proterozoic (Helikian) massifs (D_2). The changes in plunge of lineations are less likely to be due to intersecting elements of the D_0-D_3 fold sets since these possess axial trends that lie relatively close to one another. If the E-W trends are a separate event, they reflect a second post-D_2 deformation.

(B) Broad Structural Framework

Relatively detailed, but still incomplete, mapping in the Harlem Valley has demonstrated the presence of all units of the Cambrian-Ordovician shelf sequence between the bordering western and eastern pelitic highlands. As shown in fig. 2, the carbonate stratigraphy can be traced from Nellie Hill to south of Pawling, N.Y.--a distance of nearly 30 km. It is certain that continued investigation will modify fig. 2 in detail, but the larger implications

of the current map pattern are not likely to undergo substantial changes. In particular, we note that the entire valley is underlain by a complete--and overturned--section of the Wappinger Group. As shown in fig. 3, we consider this section to represent the eastern, overturned limb of a large, westward verging syncline related to the D_2 event. The axial trace of the folding is $N10^{o}$-20^{o}E. A reasonable name for this structure is the Harlem Valley Syncline. The Housatonic Massif is complementary to the syncline and appears to be a westward verging, doubly plunging anticlinorium whose overturned, lower limb is represented by the overturned carbonate sequence of the valley. Almost certainly the Helikian gneisses of the massif have been locally thrust out over the carbonate shelf sequence in the manner described by Ratcliffe (1965) for the northern Berkshires. This thrusting may be multiple and of large throw (Harwood and Zeitz, 1974). Corbin Hill may be a relict klippen of this mechanism.

Both figs 2 and 3 fail to show any complicating effects of early/late high angle faulting. As of the moment, this faulting has not been studied in detail, but it does not appear that it could markedly change the outcrop patterns as currently determined.

The regional S_2 foliation parallels the axial trace of the major overturned fold, and these two elements are taken to the genetically and temporally related. Since this foliation transects metapelites of the allochthonous Everett formation, the foliation is considered to be post-allochthonous.

The major folding and emplacement of local "hard-rock" allochthons are believed to be penecontemporaneous. The major folds and cleavage are thought to have formed during, or shortly after, the westward thrusting of the so-called "hard rock" or "High Taconic" slices. This conclusion is based upon analogy with other better understood, portions of the Taconide Zone (e.g. Zen, 1967, 1972). However, Acadian folding is known to the west (Green Pond outlier) and we must reserve the possibility that this deformation resulted in some major structures in this area (see Hall, 1968, p. 126).

When considered from a broad, regional point of view it is not difficult to envisage reasonable mechanisms leading to the formation of the Harlem Valley Syncline. Assuming a plate tectonic model similar to that of Bird and Dewey (1969) or Zen (1972), we suppose that the Middle Ordovician inversion of sea floor relief was accompanied and followed by syntectonic flysch sedimentation and the emplacement of gravity slide allochthons now exposed farther west around Pleasant Valley and Fishkill, N.Y. Continued underthrusting of oceanic crust led to increasingly severe westward directed compression that culminated in hard-rock thrust slices and the rise of Proterozoic basement units along a zone dipping to the east (fig. 2). As the basement rose from the east the overlying carbonate shelf rocks responded by overturning to the west. This overturning is most pronounced near the Proterozoic structural front. A final phase in this sequence was represented by late westward thrusting of the Proterozoic massifs (Ratcliffe, 1969). This thrusting may have been of major dimensions in Southeastern New York and Western Connecticut. The location of the Housatonic massif and the New Milford Dome within the carbonate shelf is suggestive of this.

The foregoing sequence of events provides an acceptable conceptual framework within which to understand the regional geology. However problems arise

when the geology of the valley is examined in detail. Some of these problems are discussed in a later section.

(C) Outline of Geologic History

Within the context of the foregoing regional setting, and notwithstanding some of the noted uncertainties, we suggest the following summary of events for the geologic history of the region that includes the Amenia-Pawling Valley.

Because of multiple overprints of deformation and metamorphism, portions of this history and timing are, of necessity, speculative.

(1) During Middle Proterozoic (Helikian) time a sequence of sedimentary and volcanic rocks was deposited and metamorphosed during the Grenvillian Orogeny (1100-850 mya).

(2) In Late Proterozoic (Hadrynian) time rifting of continental dimensions led to the initial opening of Iapetus (Proto-Atlancic Ocean). Eastward of the continental margin marine fault-trough deposits began to accumulate (Rensselear Graywacke, Nassau Fm.). These thick units were deposited within an age bracket of 850-570 mya.

(3a) In Early Cambrian time marine waters began to transgress the craton from southeast to northwest. This incursion is marked by the development of orthoquartzites (Poughquag Quartzite), which grade upward into the Stissing Dolostone.

Continued marine transgression resulted in the development of an extensive carbonate shelf throughout Cambrian and Early Ordovician time. This shelf is now represented by the Wappinger Group (Stockbridge Formation).

(3b) To the east of the shelf there formed a series of black shales and limestone conglomerate beds (Germantown Formation). These were followed by green shales, siltstones and cherts (Stuyvesant Falls Formation). It is believed that these units were formed on, or near, the continental slope. The presence of carbonate conglomerates and brecciolas support this contention.

(4) Near and at the close of Early Ordovician (Canadian) time, there occurred widespread high angle faulting and regional uplift. Some folding and fault block rotation may have accompanied this event (Quebecian or Penobscot Taphrogeny). The cause of the shelf breakup is not well understood, but its occurrence resulted in the discontinuous development of an Early Ordovician erosional surface on top of which residual, iron rich soils were developed. The erosional surface bevelled to all units in the Cambrian-Ordovician shelf sequence and probably to the Proterozoic basement itself. The unconformity may have extended into portions of the continental slope. Presumably the expansion of Iapetus ended at this time.

(5) As Iapetus began to diminish in size, compressional forces of the Taconic Orogeny (Bonnian Phase) resulted in a series of welts and troughs, some of which were probably off shore island arcs (Bronson Hill Anticlinorium ?). Early Normanskill pelites and bedded cherts (Indian River, Mt.

Merino) accumulated at this time and were deposited in the deeper portions of the troughs. Younger Normanskill graywackes, siltstones, and silty pelites accumulated on the slopes of the troughs. Farther to the east the island arc helped feed the eugeosynclinal sequence now represented by the Missiquoi Fm., Ammonoosuc Fm., Hartland Fm., etc.

(6) Continued compressional forces resulted in a relatively large land mass (Vermontia) during the Middle Orodvician (Mohawkian). Erosion of this landmass produced the muds, sands, and graywackes that were deposited in the trough to the west (Snake Hill-Martinsburg Trough). Presumably, the source rocks for this flysch sequence were the uplifted slope and eugeosynclinal sediments to the east. As Vermontia continued to grow, slope and basin sediments located near the axis of the uplift became gravitationally unstable and slid westward into the deepening trough (Vermontian Phase of the Taconic Orogeny). Sedimentation continued during this submarine sliding and some of the allochthonous rocks were eventually buried in younger Snake Hill-Martinsburg muds and silts. Some of these sediments may have been derived from the allochthons themselves. The emplacement of the allochthons resulted in the development of a chaotic melange in the soft pelites at the base of the slide. This mélange, or wildflysch, is well developed in western Dutchess County but has not been recognized in the metamorphic terrain of eastern Dutchess County.

(7) During the Late Middle Ordovician (Late Mohawkian) time the Snake Hill-Martinsburg Trough filled with fairly well sorted clastics of the Schenectady-Quassaic molasse. Clasts in the Quassaic conglomerates near Illinois Mt., New York denote derivation from units comprising the gravity slides--demonstrating at least partial subaerial exposure of some of these allochthons.

(8) Tectonism continued into the early Late Ordovician (Maysvillian) time and produced hard-rock thrust slices with associated carbonate and Walloomsac slivers torn from the older, subjacent shelf. In Dutchess County these slices are represented by the Everett Schist and by some plates of Proterozoic (Helikian) gneiss. Accompanying, or immediately following, the hard-rock slices there developed westwardly overturned folds and regional development of cleavage. Mineral ages of ∿400 mya suggest that a pulse of regional metamorphism occurred at this time (Long, 1962). These ages are most prevalent in western Dutchess County but appear to have been overprinted in eastern Dutchess County.

(9) During the Late Ordovician (Richmondian) mafic igneous bodies were emplaced at Cortland and Bedford (∿435 mya, Long, 1962).

(10) During the latest Ordovician (Gamachian) and early Silurian (Llandoverian, Wenlockian) there occurred a widespread episode of normal, block-faulting. This is particularly well displayed in the Mohawk and Champlain Valleys and the faults are observed to cut Taconian thrust sheets. Evidence strongly suggests that these Silurian faults were accommodated along reactivated Proterozoic basement fractures. Uplands produced by this post-Taconian block-faulting provided erosional debris for the Shawangunk-Vernon-Bloomsburg clastics.

(11) During Late Silurian (Ludlovian) time evaporite deposits accumulated in central New York. Corresponding events in eastern New York and westernmost New England are uncertain. Some renewed compression may have occurred.

(12) In the latest Silurian (Pridolian) and Early Devonian (Helderbergian), crustal stability prevailed with attendant carbonate and reef development. Uplift followed, but the nature of this is uncertain. The succeeding Oriskany sands and Esopus-Carlisle Center silts and pelites suggest renewed deformation in eastern New York (Phase I of Acadian Orogeny). Brief crustal stability with Onondaga carbonates and reefs ensued. The intense Phase II of the Acadian Orogeny followed with westward overturned folding and with probable high-angle reverse faulting and metamorphism in easternmost New York. East of Wappinger Creek Valley earlier Taconian cleavage was folded. Vigorous erosion of uplifted land created the thick and extensive Catskill clastic wedge during the Middle Devonian (Erian) and early Late Devonian (Senecan). By late Late Devonian (Chautauquan) time, the Acadian Orogeny was over.

(13) The effects of Late Paleozoic deformation, (if present) in eastern New York are vague. A thermal event of about 250 mya is known in western Connecticut and it is reasonable to assume that its presence was felt in southeastern New York.

Major Problems of Local Interest

For the moment, at least, the most severe problems in the area are:

(1) The angular relationship between the Balmville Limestone-Walloomsac Schist and the inverted Wappinger units below the early Middle Ordovician unconformity. Related to this are implications concerning the nature of the pre-Walloomsac, D_0, event.

(2) The subdivision and correlation of the Manhattan Schist, that forms the eastern wall of the Amenia-Pawling Valley from south of Pawling to the Wingdale-Bull's Bridge gap. Stratigraphic assignment will help determine whether these schists are allochthonous, autochthonous, or parautochthonous.

(3) The nature of the basement rocks underlying the schist mass referred to in (2)--i.e. is the schist directly underlain by Proterozoic gneiss or Wappinger carbonate units?

(4) The structural relationships, and origin, of the isolated mass of Proterozoic gneiss exposed at Corbin Hill.

(5) The relationship of the Harlem Valley to the regional setting comprising the various Precambrian massifs of the area; the presence of Wappinger carbonates <u>east</u> of Precambrian gneisses; and the ever problematical Cameron's Line. Unravelling of the regional geology in southeastern New York and Western Connecticut represents a fundamental key to the understanding of the evolution of the Appalachians. We shall not pursue this major undertaking within this report.

Discussion of Problems

In what follows we will briefly discuss problems (1) - (3). Problem (4) (Corbin Hill) is treated in the text for Stop 6. Problem (5) must await further research.

1. Middle Ordovician Unconformity

A Middle Ordovician unconformity is widely recognized throughout eastern North America (Rodgers, 1970). Most workers have agreed that the unconformity developed following the Early Ordovician (Canadian) breakup of the shelf sequence by N-S block faulting. The faulting appears to step up the Precambrian basement near the eastern margin of the shelf and represents events immediately preceding the emergence of Vermontia and the deepening of the Snake Hill-Martinsburg Trough. Just prior to this deepening an erosional surface formed on subaerially exposed blocks. Subsequently, Balmville Limestone and Walloomsac flysch were deposited above the locally developed erosional surface. Presumably sedimentation continued unimpeded in non-exposed basins (grabens).

The foregoing exposition of the pre-Walloomsac tectonism (Penobscot or Quebecian Taphrogeny) has much to recommend it. Zen (1968), Thompson (1959), as well as numerous others, have shown that block faulting was the most probable mode of deformation at this time. However, local compressional events have been recognized and Thompson (1959), Zen (1961, 1967, 1968), and Ratcliffe (1969) have demonstrated that reality of pre-Walloomsac folding within the shelf sequence. Neumann and Rankin (1966), Ayrton (1967), and Hall (1969) have demonstrated strong compressional events of pre-Walloomsac age in Penobscot County, Maine; the Gaspe Penninsula; and the Notre-Dame-Sutton Mt. Anticlinoria. It appears as if compressional tectonics were more intense in off-shelf than in on-shelf environments.

The importance of understanding the pre-Walloomsac event is emphasized when dealing with overturned Wappinger carbonates. Thus, Ratcliffe (1969a, p. 2-12) was able to demonstrate at No Bottom Pond Window in the State Line Quadrangle in eastern Columbia County, N.Y., that an angular discordance of 70° existed between the Balmville Limestone and the underlying Stockbridge carbonates. Because of the detailed field evidence, Ratcliffe concluded that the Stockbridge had been folded, and overturned, prior to deposition of the Balmville Limestone.

Within the Amenia-Pawling Valley the Balmville-Walloomsac sequence is found in patchy exposures lying above overturned Wappinger carbonates. It appears from field relationships that this situation need not imply inversion of the Wappinger Group prior to deposition of Balmville-Walloomsac units. It is equally possible that gentle folding and even tilted block faulting, could have provided a westwardly dipping Wappinger section which was bevelled down and then overlain by the younger lithologies (fig. 4). Subsequent overturned folding could have brought the units into their present configuration. The preservation of patches of Balmville and Walloomsac would be enhanced if the major folding episode that overturned the Wappinger Group deformed the Middle Ordovician unconformity also. In so doing there could have resulted

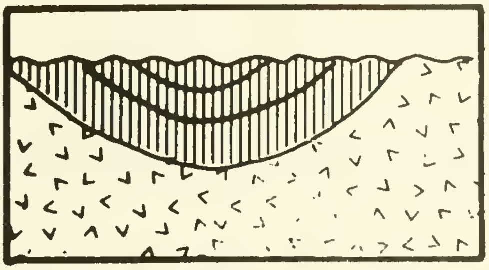

Fig. 4a - Development of early Middle Ordovician unconformity above gently folded shelf.

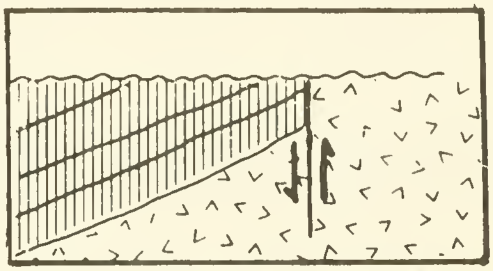

Fig. 4b - Development of early Middle Ordovician unconformity above rotated fault blocks.

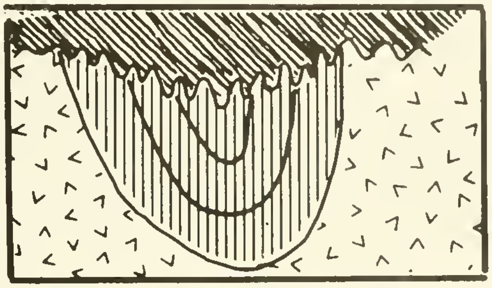

Fig. 4c - Middle Ordovician folding following deposition of Balmville-Walloomsac. Note folding of unconformity.

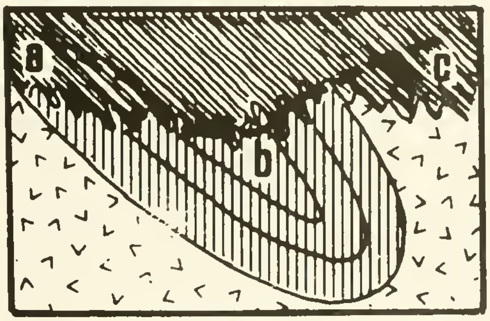

Fig. 4d - Culmination of Middle Ordovician folding and overturning. Unconformity gets in-folded. The dashed line abc represents a present day erosion surface.

(Symbols as in Fig. 2)

in-folded keels of Balmville-Walloomsac which escaped later erosion (fig. 4). Only detailed field investigations of the angular relationships on either side of the unconformity can resolve this situation. Unfortunately critical outcrops are lacking. In the absence of information to the contrary, we choose to regard the local Pre-Walloomsac event as primarily non-compressional, and we attribute the formation of the overturned folds to later Taconian or Acadian events.

As a final observation of this matter, we note that the possibility exists that, locally, the Balmville and Walloomsac were thrust into their present positions relative to the overturned carbonate sequence. This possibility appears ad hoc and is not favored.

2. Subdivision and Correlation of the Manhattan Formation (?)

The problem of subdividing and correlating Taconide pelitic masses constitutes one of the historical pivot points in the time honored Taconic controversy. The difficulties inherent in this undertaking are complicated by high metamorphic grade. In the western gravity slides fossil control, color differences, textural differences, bedding characteristics, etc., have been helpful in providing stratigraphic control. However, these criteria are not generally present in the later, hard-rock slices lying to the east. Here the stratigraphic divisions have usually been reduced to the recognition of two major units: the autochthonous Walloomsac Fm. and allochthonous slices which, in much of Dutchess County, N.Y., have been referred to as the Everett Schist (Hobbs, 1893). The distinction between Everett Schist and Walloomsac phyllites is not usually obvious. Often, the Walloomsac is darker, rustier, and more graphitic than the Everett; the latter tending to have a greenish or silvery hue. As metamorphic grade increases, these distinctions become less obvious.

The 15 mile long ridge that defines the eastern margins of the Amenia-Pawling Valley from south of Pawling to the Wingdale-Bull's Bridge gap is underlain for 5-6 miles to the east by enigmatic schists of the type described in the preceding paragraph. On the 1973 edition of the New York State map these are shown as Manhattan Fm., and we have retained this nomenclature for the purposes of this field guide. For the most part these rocks consist of coarsely micaceous sillimanite-staurolite-garnet-muscovite-quartzo-feldspathic schists. They closely resemble units mapped as hard-rock slices of Everett Schist on the northwestern side of the carbonate valley (see Stop 3). The strongest argument for correlation of these rocks with the Everett rests with their lithologic similarity to high grade Everett in other areas. In general, workers have tended to regard the Everett as more aluminous than the Walloomsac, and this difference is reflected in a greater ratio of muscovite to biotite in the former. In addition the Everett generally displayes more staurolite, chloritoid, and alumino-silicates than does the Walloomsac. These criteria suggest that the rocks in question should be assigned to the Everett rather than to the dark, rusty weathering, graphitic Walloomsac. However, this assignment rests on no quantitative, or unequivocal, evidence.

The distinction between Walloomsac and Everett is analogous to that between the Lower and Upper Manhattan (Manhattan A and C of Hall, 1969) as

reported by Ratcliffe and Knowles (1969), and discussed in the straigraphy section of this report. While it is not at all certain that the main mass of the schist in question here is correlative with Manhattan C, the litholigic similarities are pronounced, and we adopt this correlation as a preferred alternative. Hall (1968) and Ratcliffe and Knowles (1969) have suggested that the Manhattan C may be allochthonous. We suggest here that a similar possibility exists for the schists underlying the matapelite ridges of Pawling, New York.

Evidence favoring an autochthonous history for the eastern schist mass derives principally from the presence of Balmville Limestone and dark, rusty, and calcitic Walloomsac schists underlying the main schist mass at its northern margin in the Wingdale-Bull's Bridge gap (Balk, 1936; Waldbaum, 1960). However, this data is in no way inconsistent with the general Taconic situation in which allochthonous masses overroad the black shales of the shelf.

At the base of the schist mass directly east of Pawling the basal Walloomsac and Balmville are not present and the coarse, muscovite rich schists lie directly upon Stissing Dolostone and even Poughquag Quartzite. Inspection of the map in fig. 3 shows that the schist transects the Stissing contact and even cuts the Stissing out entirely near the southern end of the valley. The fact that the Stissing is here overturned is further suggestive of a tectonic contact, but such a contact could be the result of the early Middle Ordovician unconformity, or of local westward thrusting of Walloomsac, and need not indicate a far travelled hard-rock slice of Everett.

While we prefer an Everett assignment, and an allochthonous history, for these rocks, we re-emphasize that the matter remains equivocal.

3. Nature of the rocks underlying the Manhattan Schist

It is not possible to know with any certainty whether the mass of Manhattan Schist is underlain by units of the carbonate shelf or by Proterozoic gneisses. It is conceivable that beneath the Manhattan there exists an eastward dipping, right-side-up sequence of carbonates that represent the eastern limb of the southern extension of the Housatonic Anticlinorium. It is equally likely that the schists are underlain, at least in part, by Proterozoic gneisses near the axis of the anticlinorium. This possibility is favored by the fact that the western margin of the schist overlaps the Stissing Dolostone, and the Proterozoic basement can be at no great depth. Furthermore, the schists are surrounded on their southern and southeastern margin by Proterozoic gneisses of the Hudson Highlands.

One reason for preferring at least a partial Proterozoic gneiss sequence below the Manhattan Schist is that its presence provides a reasonable source area for the Proterozoic gneiss outlier at Corbin Hill (see Stop 6).

Summary

Although important unresolved problems remain in the Amenia-Pawling Valley, we are able to conclude with reasonable certainty that the valley

is underlain by the overturned, eastern limb of a large NNE trending syncline. In the western limb of this fold the carbonate shelf remains hidden beneath Walloomsac and Everett schists. This structure is termed the Harlem Valley Syncline and is thought to be complementary to the Housatonic Highland massif and its southern extension. It is suggested that the Proterozoic gneisses at Corbin Hill are a relict klippen of a hard-rock thrust slice emplaced in the late Taconian Orogeny (Hudson Valley phase). Local allochthons of Everett Schist were emplaced at the same time. Penetrative cleavage and metamorphism followed these events, probably, in late Ordovician time. A Devonian overprinting of Ordovician metamorphism is probable.

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Road Log

Mileage	
0	Intersection of Mass 41 and US 7 in Great Barrington. Take Mass 41 south.
1.5	Intersection of Mass 41 and Mass 71. Continue south on Mass 41.
3.9	Intersection of Mass 41 and Mass 23. Continue south on Mass 41. Carbonate exposure at intersection.
8.3	Everett Mt. in N-S ridge to the west of Mass 41. This is the type locality of the Everett Schist.
11.7	Mass-Conn State Line. Carbonates on west side of road.
16.6	Junction of Conn 41 and US 44 in Salisbury. Continue SW on 41-44.
18.3	Junction US 44 and Conn 41 in Lakeville. Continue south on Rt. 41.
20.0	Junction with Conn 112.
20.3	Briarcliff Dolostone on east side of road.
20.7	Briarcliff Dolostone on both sides of Rt. 41.
25.3	Junction with US 4 in Sharon. Continue west on Rt. 41.
25.5	Carbonate on south side of road.
25.8	Carbonate on north side of road at Filston House.
27	Walloomsac black slates on south side of road.
28.1	Walloomsac black slates on north side of road.
28.3	Carbonates on hill to north of road.
30.3	Intersection of NY 22, NY 343, and US 44 in Amenia, N.Y. Turn south on NY 22.
31.1	Walloomsac schist and upper Wappinger carbonates (Copake ?) on NW side of road.
31.7	Stop 1. Roadcuts in Walloomsac schist and Wappinger carbonates on both sides of NY 22. Here the Walloomsac schist overlies calcitic and dolomitic limestones of the upper Wappinger--possibly the Copake Limestone. The presence of brown and tan weathering dolostone beds in the carbonates preclude their being Balmville. Probably this was a site of non-deposition of the Balmville. At or near the contact, there is developed tectonic interleaving of schist and carbonate. This interleaving could result from at least two causes: (1) differing mechanical properties of the schists relative to the carbonates;

(2) the shearing off of original irregularities along the old erosional surface. Within the outcrop, the foliation and bedding surfaces strike N40E and dip 50^{o}W. The stratigraphy is right side up. Tight folding takes place about N10-15E axes and the folds verge westward. Near the south end of the outscrop there exist good examples of the relationship of fold wavelength to bed thickness. Note that the principle foliation has been refolded representing, probably, the D_2 and D_3 events. A thin section of Walloomsac from this outcrop shows abundant quartz and biotite as well as garnet, feldspar, muscovite, graphite, and metallic opaques. Two generations of biotite can be seen megascopically.

37.6 State Police headquarters in Dover Plains, N.Y.

39.1 Stop 2. Nellie Hill. Large roadcuts in upper Wappinger carbonates. The beds strike N10-20E and dip 30^{o}-40^{o}E. In the fields beyond the west side of the road are outcrops of Balmville Limestone and Walloomsac Schist that also dip to the east. If one proceeds eastward over the top of the roadcut and onto the next ridge ($\sim$300 m), Briarcliff Dolostone is encountered. Still further to the east the Pine Plains Formation, Stissing Dolostone, and Poughquag Quartzite crop out successively until the Proterozoic gneisses of the Housatonic Highlands rise in the ridge defining East Mt.. Except for minor folding, the strata dip consistently to the east and the entire section must be regarded as overturned. Chestnut Ridge, immediately to the west of Rt. 22, consists of Everett Schist and is regarded as an allochtonous hard-rock slice.

The southern portion of the outcrop is thought to consist of Copake Limestone that has been pervasively folded about N10-20E axes, plunging 10^{o}-15^{o} N. Excellent examples of transposed bedding and sheared out fold limbs can be seen. The rock consists of dark, pure calcitic metacarbonate interbedded with coarser sandy doloston Possible crossbedding can be seen in the steep walls of the roadcut

At its northern end, the outcrop shows the development of well layered buff and brown dolostone beds interlayered with dark, massive calcite rich beds. These units are thought to belong to the upper portion of the Rochdale Limestone. Bedding averages around 0.5-1 m in thickness. Some of the dolostone beds are quartzose anc show thin bedding laminations. No cross-bedding or graded bedding has been recognized and discoveries of the same will be welcomed.

Above the road level, and within the tree cover, there are developed coarse, gray, massive limestones that extend to the top of the hill. These are considered to be part of the Rochdale. Betweei the hilltop and the Briarcliff dolostone, limestone bearing units of possible Halcyon Lake assignation crop out.

A problem with the correlations as given above is that the resulting thickness of the Rochdale Limestone is less than would be expected. Warthin (pers. comm.) reports approximately 125 m of Rochdale near Poughkeepsie. If the Halcyon Lake is present in the section here, it does not appear possible to have 125 m of Rochdale.

Perhaps faulting has cut out some of the section. Alternatively the Halcyon Lake Fm. may not have been deposited locally. A further possibility is that the beds assigned here to the Copake are actually Rochdale.

39.8 Stop 3. Park on east side of NY 22 near bend in road. Walk southward along railroad tracks, for approximately 150 meters. Excellent outcrops of Everett Schist are exposed in a small railroad cut. NO HAMMERS PLEASE. Large (1/8" - 1/4") staurolite and garnet crystals are developed in coarsely micaceous muscovite schists which display the typical silver sheen of the Everett at this grade. Quartz and feldspar are plentiful with quartz predominating. Some graphite is present. The foliation has been refolded.

40.5 Briarcliff Dolostone in roadcut.

40.6 Briarcliff Dolostone in roadcut.

41.1 Briarcliff Dolostone on hill to west of road.

42.3 Briarcliff Dolostone on east side of road.

42.5 Briarcliff Dolostone on east side of road.

42.6 Leave NY 22 and turn east on Crickett Hill Road (unmarked).

42.8 Abandoned quarry in Briarcliff Dolostone to south of road.

43.0 Cut in Briarcliff Dolostone.

43.3 Cut in Briarcliff Dolostone.

43.6 Cut in Pine Plains Formation.

44.7 Junction with NY 55. Proceed east on Rt. 55.

45.8 Turn north on Dutchess County 22 and cross bridge over Ten-mile River.

46.1 Manhattan C or Waramaug Schist on north side or road.

46.7 Housatonic Highlands directly ahead.

47.1 Poughquag Quartzite along east side of road.

47.3 Dogtail Corners. Continue directly across junction and onto Preston Mt. Road.

47.8 Stop 4. Exposures of Poughquag Quartzite and Proterozoic (Helikian) gneisses of the Housatonic Highlands. The contact between gneisses and Poughquag has probably been faulted and a small valley separates them. The Poughquag shows its typical white to brown and pink coloration. Small quantities of mica are present on some foliation surfaces. Turn cars around and head back to Dogtail Corners.

49.3 Junction of Dogtail Corners. Turn west.

49.7 Turn west on small dirt road.

49.9 Outcrops of Stissing Dolostone.

50.1 View ahead of Peckham Industries quarry in Stissing Dolostone.

50.5 Ledges of Stissing on south side of road.

50.7 Turn south at T-intersection.

51.0 Entrance to Peckham Industries Quarry. Park cars in quarry yard. Stop 5. The quarry is within Stissing Dolostone and the vast expanse of dazzling white dolostone provides insight into the purity of this lowermost carbonate. During World War II, this quarry was utilized as a source for magnesium. Company officials have provided analyses of the dolomite, and these will be discussed at the meeting. Within most units the rock consists almost entirely of dolomite and calcite. The structure within the quarry seems to be fairly straightforward. Bedding dips steeply around an anticline that trends N10-20E and plunges 10^{o}-15^{o}S. The core of this anticline is preserved in the lower quarry level where pelitic beds of Stissing are also exposed. Presumably these are related to the red shale horizons recognized by Mrs. Knopf (1946) near the middle of the Stissing. At the stratigraphic level of the pelite rich zones, the Stissing is difficult to distinguish from portions of the Pine Plains Formation.

In terms of regional structure, note that the quarry lies within the Wingdale-Bulls Bridge gap. Within this gap the Stissing Dolostone appears to wrap around the southern end of the Housatonic Highlands. Moreover, Stissing within the gap appears to be structurally and stratigraphically continuous with Stissing to the west in the Harlem Valley. This strongly suggests that the gap represents the south plunging nose, and upper limb, of the westward verging anticlinorium cored by the Proterozoic gneisses of the Housatonic Highlands. The possibility exists that this anticlinorium is developed on an eastward dipping thrust plate (Harwood and Zeitz, 1974).

Return to cars. Leave quarry and turn south at entrance.

51.9 Cross Ten-mile River. Turn south on west side of bridge.

52.5 Intersection with NY 55 at Webatuck.

53.3 Intersection of NY 55 and NY 22. Turn south on NY 22.

53.9 Stop light at Harlem Valley State Hospital.

54.6 Pine Plains Formation on west side of Rt. 22.

55.0 Pine Plains Formation on east side of Rt. 22.

55.3 Manhattan Formation of the schist mass comprising the east side of the valley.

55.6 Pine Plains Formation on west side of Rt. 22.

55.8 Pine Plains Formation on west side of Rt. 22.

56.1 Stissing Dolostone on both sides of Rt. 22.

56.5 Stop 6. The long roadcuts on either side of NY 22 are fine examples of Pine Plains Formation. However we will not examine these at this location. The primary purpose for this stop is to point out, and discuss Corbin Hill which rises out of the swampy fields to the west of NY 22.

Corbin Hill consists of Proterozoic (Helikian) gneisses whose bedding and foliation are conformable to the valley trends. Balk, 1936, considered Corbin Hill to represent a slice of basement brought to its present erosional level along a steeply dipping reverse fault block that involved Precambrian rocks only (fig. 5a). If this mechanism is correct, then it should be reflected by a break in the stratigraphic succession of the valley carbonates. Similarly, if Corbin Hill punched its way upward as an elongate domal mass, then the carbonate stratigraphy should wrap around the Precambrian gneisses (fig. 5c).

Mapping around Corbin Hill has shown that the carbonate stratigraphy appears to be unaffected by the gneiss body. Units of the Wappinger Group can be followed down the valley and "through" Corbin Hill with no signs of displacement of "wrapping-around". As shown in fig. 2, the Briarcliff Dolostone appears to underly Corbin Hill. The presence of upper, calcitic units of the Wappinger lying just beyond the northern terminus of Corbin Hill is inconsistent with a steeply upthrust, or up-punched, origin this feature. Similarly the presence of Balmville Limestone, Walloomsac Schist, and even Everett Schist in close proximity to the western margin of the Precambrian gneisses makes it difficult to account for Corbin Hill by essentially autochthonous mechanisms. The stratigraphic continuity of the carbonate shelf units, and their overturned attitudes, cannot tolerate autochthon-based models for Corbin Hill. In effect, the gneisses on Corbin Hill do not appear to be rooted through the valley carbonates. While this conclusion cannot be made firm without further evidence (largely geophysical) it does represent our preferred interpretation of the field data.

In the absence of further data, we are unable to offer any convincing, detailed history relating to the origin of Corbin Hill. Our preferred model is that these gneisses represent an erosional outlier of a low dipping hard-rock slice that transported Proterozoic rocks from east to west. Rather than suppose the existence of extensive Proterozoic rocks in the slice (fig. 5d), we prefer to attribute the rocks of Corbin Hill to tectonic slivering of basement rock by an overriding slice of Everett Schist (fig. 5e). This slivering could have occurred in the basement rocks that underly the Manhattan Schist terrain, forming the ridge on the east side of the valley. While other source areas exist, this one appears to be the most economical of long distance transport. Similarly, slivering of the Proterozoic offers the simplest explanation for the restricted outcrop of Corbin Hill.

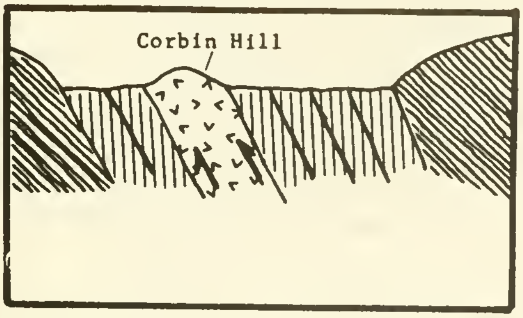

Fig. 5a - Corbin Hill as an upthrust block (Balk, 1936). Inconsistent with stratigraphy.

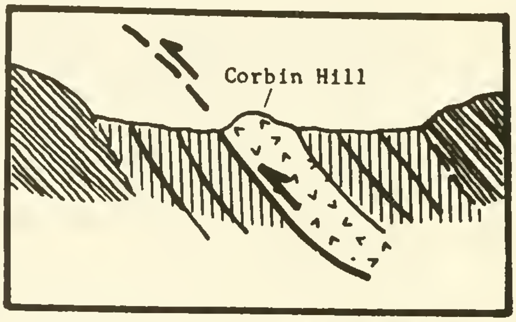

Fig. 5b - Corbin Hill as the basal unit in a large thrust sheet. Inconsistent with stratigraphy.

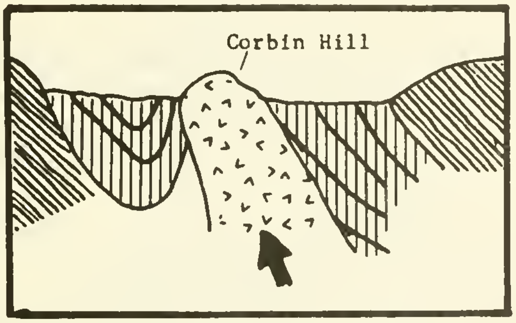

Fig. 5c - Corbin Hill as an up-punched gneiss dome. Inconsistent with stratigraphy.

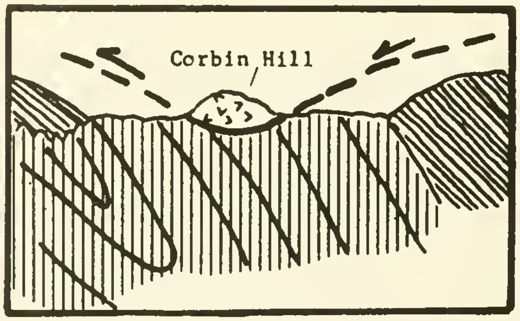

Fig. 5d - Corbin Hill as a far traveled thrust slice.

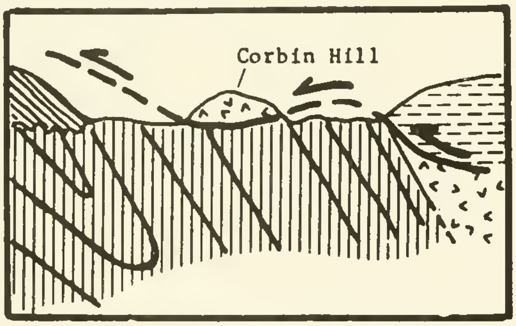

Fig. 5e - Corbin Hill as a tectonic slice on the sole of an Everett thrust sheet.

(symbols as in Fig. 2)

The foregoing model is consistent with the recent aeromagnetic interpretations of Harwood and Zeitz (1974) for rocks of the Housatonic massif. Here, eastern, weakly magnetic Precambrian rocks are thrust westward along low angle faults rooting in the east. This thrusting occurred late in the Taconian Orogeny and involved the various hard-rock slices of the High Taconics and the Precambrian massifs. Note that just to the north of Towners, New York, Balk (1936) mapped Precambrian and Poughquag Quartzite thrust over carbonate units. Lying north of this is the small Pine Island mass of Precambrian and Poughquag which, presumably, is also thrust in (fig. 3).

56.9 Pine Plains Formation on east side of NY 22.

58.6 Trinity-Pawling School.

59.2 Briarcliff Dolostone on east side of NY 22.

59.6 Briarcliff Dolostone on east side of NY 22.

59.9 Signal. Briarcliff Dolostone in large roadcut on east side of NY 22. Abundant diopside and tremolite are developed in the outcrop.

60.2 Pass under NY 55.

60.3 Briarcliff Dolostone on west side of NY 22.

60.9 Slow down and turn left across divider. Head back north.

70.0 <u>Stop 7</u>. Road cut in the Pine Plains Formation. Excellent example of the highly variable lithologies that characterize this unit. Brown and buff sandy dolostones alternate with quartzites and relatively pure, massive white dolostones. Tan colored units often shows typical rotten weathering. Punky, asphalt bearing layers give off H_2S upon breaking open. Bedding is of variable thickness. The beds of quartzite and chert have undergone boudinage and pinch and swell of textbook quality. Reaction rims and selvages exist between the carbonates and the quartz rich layers. In the brown to purplish pelitic zones phlogopite, sphene, diopsidic pyroxene, and tremolite are developed. Within the more massive beds of grey weathering, white colored dolostones diopside tablets attain dimensions approaching 3 cm across.

On the east side of NY 22, the Pine Plains units strike subparallel to the road and dip steeply to the east. On the west side the strike has turned E-W and dips are steeply to the south. This represents a fairly open, dextral type of fold that swings the carbonate units and the Manhattan Schist westward for about 0.8 km at which point strikes return to NNE trends (see fig. 3).

The contact between the Briarcliff dolostones and the Pine Plains Formation is thought to occur just to the west of NY 22. The low hill rising from NY 22 is known to be underlain by Briarcliff Dolostone and this unit is beautifully developed just to the south of

the NY 55 underpass (0.5 km along strike to the north). Perhaps some of the massive, pure dolostone at the north end of the cut should be assigned to the Briarcliff.

Minor folds in the outcrop suggest that we may be observing here the limbs of larger isoclinal folds. Two foliations exist and are best manifested by micaceous bands.

70.7 Turn onto entrance ramp for NY 55 west.

71.1 Briarcliff Dolostone with excellent diopside crystals on north side of NY 55.

71.7 Stop 8. Very large roadcuts in the Briarcliff Dolostone. The Briarcliff consists typically of grey weathering, light colored rather pure dolostones with yellow to shite and even black chert layers (1"-2") abundantly developed in some units. Knots and nodules of vitreous quartz are locally present and weather out above the dolostone surface. At the east end of the cut some dirty portions of the Briarcliff exhibit moderate development of phlogopite. At the western end of the beds of dolostone are massive and pure. This difference appears to be reflected in the more open style of folding associated with the pure thick layered (5-7 m) beds.

A large number of different structural styles and phenomena can be seen in the roadcut. Folds range from fairly open flexural styles to isoclinal folds that may involve flowage and/or significant flattening. In many areas of the cut disharmonic folding is pronounced with the dolostones undergoing extensive flowage while the much more brittle chert layers show rupturing and extensive separation of blocks. Examples of the folded boudinage are beautifully developed at the eastern end of the roadcut.

Judging by fold style, and intensity, there appears to be at least two major compressional events recorded in the outcrop. The first is represented by many of the tight folds. Tracing out of beds suggests that early axial planes have been folded. An excellent example of a refolded isoclinal fold can be seen on the south side of NY 55 about 50' west of a pronounced saddle in the roadcut. Plots of linear elements shows a pronounced maximum at N10-20E, 10^{o}NE and a scattering of other lineations in directions ranging from NE to SW. Plunges of fold axes are variable. These relationships are not well understood with the context of the D_0-D_3 events described earlier. Perhaps D_2 and D_3 were close to coaxial in this region.

Numerous high angle faults cause observable offsets in the dolostones. Some of the fault stones contain serpentine. These faults appear related to a larger fault zone that causes a topographic saddle about half way along the roadcut.

Of particular petrologic interest are layers of tremolite and diopside in the Briarcliff. These are best observed on the top of the roadcut at its southeastern end. Massive beds of tremolite and diopside areas are mutually exlusive and seem to reflect the

relative immobility of the vapor phase during metamorphism. Note that the diopsides, especially, appear to post-date any severe orogenic events.

72.6 Calcitic Walloomsac schists containing calc-silicates, calcite, and overlain by a calc-silicate bearing calcitic dolostone. The latter may be a tectonic sliver similar to those seen at Stop 1.

72.8 Stop 9. In this small roadcut we are afforded a view of the contact between the Balmville Limestone and the Walloomsac Schists. At the north end of the outcrop the schists overly the Balmville, but at the south end, units dip steeply to the east, and, if outcrop were preserved, the Balmville would overly the schists. We consider this relationship to be due to proximity to the hinge line of the overturned Harlem Valley Syncline whose upright, western limb is entered as NY 55 is followed to the west (see fig. 3). The roadcut itself probably represents a minor fold near the hinge line since black Walloomsac calcitic schists are found farther to the east at mileage 72.6.

End Road Log

Trip C-11

Polyphase Deformation in the Metamorphosed Paleozoic Rocks East of the Berkshire Massif

by

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Introduction

The purpose of this trip is to demonstrate the following; a) the age relationship and geographic distribution of four generations of minor folds, b) the relationship of each generation to the major structures outlined by detailed stratigraphic mapping, and c) the structural evolution of a segment of the eugeoclinal belt east of the Berkshire Massif. The evolution begins with intensive thrusting in the Taconic orogeny (Middle to Late Ordovician) culminates with multiple generations of regional folds in the Acadian orogeny (Middle to Late Devonian) and terminates with normal faulting of Triassic age.

The Cambrian and Ordovician stratigraphy of the area consists of metamorphosed (kyanite-sillimanite grade) shales, graywackes, mafic volcanic rocks and minor amounts of chert and serpentinite. Cyclical bedding, some of which is graded, is common in the Cobble Mountain Formation of Middle Ordovician age but is rare in the older part of the section below the Moretown Formation. Abundant feldspar in the gneiss of the upper member of the Cobble Mountain Formation suggests that this unit is a flysch deposit derived from the island arc complex now represented by the rocks along the Bronson Hill anticlinorium to the east. A complete description and regional correlation of the Cambrian and Ordovician section is discussed in Hatch and Stanley (1974).

The regional setting of the Blandford-Woronoco area is discussed in other trips in this volume and need not be repeated here. Interested readers are referred to Stanley (1975) and other papers included in U.S. Geological Survey Professional Paper 888.

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Method of Study

In order to understand the axial surface maps of each fold generation (figs. 1, 3, and 5) it is necessary to describe in some detail the methods employed in this study. Minor folds are particularly useful in revealing the finite strain of deformation, the geometry and age of major structures, and structural chronology. Before any of this can be done, however, it is essential to separate folds into coeval systems. In this study minor fold generations are designated, from oldest to youngest, F_1, F_2, ... F_n. Each generation is considered the result of a broadly coeval stress field.

Within any one area of contiguous outcrop, the age sequence of minor folds was based on the principle of superposition -- a fold was considered younger if it deformed the axial surface and, in some localities, the axes of an older set or sets of folds. After the age sequence was established in well-exposed localities, the style, orientation, and intensity of deformation for each fold generation were documented for each rock type. Data on the attitude of hinge and axial surface were recorded for a number of folds of each generation so as to evaluate the overall consistency, or lack thereof, in orientation. Although quantitative measures of strain and deformation were not used in this work, qualitative estimates were based on fold tightness and cleavage development; the latter criterion proved to be the most meaningful parameter of deformation intensity throughout the areas of study. Most of my work was centered in the Cobble Mountain Formation, because this formation in the Blandford-Woronoco area contains the most complete and best preserved sequence of superposed folds. Descriptive parameters of each generation are summarized in table 2 (Stanley, 1975).

Correlation of minor folds between localities of extensive outcrop within a quadrangle was based on relative age, style, and orientation. Although the form of the axial-surface cleavage proved to be the most reliable characteristic of each fold generation, no single criterion was consistently dependable. The age criteria are most successfully employed where the distance between outcrops is small; they become less reliable where the distance is 16 or 24 km (10 or 15 miles), at which scale orientation is usually of little value, and style and relative age can only be used with considerable caution. Any scheme of regional correlation must be evaluated by the density and aereal distribution of reliable data on superposition. Correlation of structural

events among separate areas within a region can be based on the geological age of the rocks in which a given fold generation is developed and on the geometric relationship between minor and major folds. If, regarding the latter parameter, the statistical axial surface of a given generation of minor folds is parallel to the calculated axial surface of a major fold, then the two are considered coeval. Regional correlation of major structures implies that their associated minor folds are correlative in a structural sense. Using these methods, I have been able to assign and correlate approximately 2,000 minor folds to their proper fold generation within and among the areas in western Massachusetts and Connecticut (Stanley, 1975).

Field data on each minor fold generation were analyzed in spherical projection and on geologic maps (figs. 1-6). Equal-area nets of fold hinges and poles to axial surfaces were prepared for subareas within each area. Subareas were delineated on the basis of the pattern of axial-surface poles which formed either diffuse point maxima or fairly complete great-circle girdles (figs. 2, 4, 6). Geologic maps showing axial surfaces of minor folds were prepared for each fold generation (figs. 1, 3, 5). Axial surfaces were used instead of fold axes because: (1) they proved to be the most reliable age criterion; (2) they provide a qualitative estimate of the intensity of deformation; (3) they record the presence of younger fold generations because they are essentially planar before they are subsequently deformed; and (4) they represent a principal symmetry plane of finite strain (perpendicular to $\lambda 3$, axis of minimum quadratic elongation).

The axial-surface maps were constructed by plotting all available orientation data for a given minor fold generation. The orientation of axial surfaces between data points was interpolated where the surrounding coverage was dense. The density of these surfaces on each map represents the abundance of data in any one area. The resulting maps (figs. 1, 3, and 5) show: (1) the systematic change in orientation of axial surfaces from place to place; (2) the areas of intense deformation associated with each fold generation; and (3) the relationship of each minor fold generation to the major structures of the area.

Fold Generations - discussion

The minor folds in the Blandford-Woronoco area are grouped into generations F1 to F4 according to superposed relationships displayed in such well exposed areas as the southeastern part of Cobble Mountain Reservoir (Stops 1-8, fig. 8).

Of the eight style characteristics listed in table 1 (Stanley, 1975), profile form and axial surface cleavage are by far the most useful and reliable identifying features of each generation. Folds of the two older generations are commonly tight to isoclinal in profile. An excellent axial surface schistosity is everywhere present in F2 folds. Although F1 folds nowhere have a cleavage marking the axial surface, many of the F2 hinges contain micas oriented at an angle to the F2 axial surface schistosity and the bedding. These micas form limbs of very tight chevron folds whose axial surfaces are parallel to the F2 schistosity. The limb micas represent an older schistosity that has been almost totally obliterated by the development of the penetrative schistosity of the F2 folds. Examples of F1 folds will be seen at stops 1-3 (fig. 9).

Folds of the two younger generations are generally crenulate in profile in the schists of the area and are commonly more open and quite distinct from the older two generations. Cleavage is commonly not present in the youngest fold generation (F4), except for a crenulate cleavage that is poorly developed in some of the tighter, more highly deformed folds of this group. F4 folds are most easily recognized where the crenulate cleavage, slip cleavage, or spaced schistosity of F3 folds is systematically bent by F4 (stop 7, fig. 9).

Folds of generation F3 display the greatest range in cleavage development. From stop 1 to stop 7 in the Cobble Mountain area very weakly developed crenulate folds without cleavage are progressively deformed to folds first with crenulate cleavage,1/ then with slip cleavage, and finally with a well-developed spaced schistosity.

These stages represent higher levels of strain within F3 and are also used as a indicator of the intensity of deformation for other fold generations. It must be emphasized, however, that the strain represented by each generation is a summation of strain associated with the development of that generation plus the additional strain imposed on it by younger superimposed folds. No attempt has been made to separate the two in this study.

Let us now consider the intensity of strain and the range of intensities associated with each fold generation as indicated by cleavage. Little can be said for F1 because the number of observations is few. A schistosity presumable existed but it has been largely obliterated by recrystallization and repeated deformation in the three younger fold generations. F2 folds represent a uniformly high level of strain, because the associated cleavage is everywhere a well-developed, penetrative schistosity. The strain level

of F3 shows the treatest range of all four generations. The strain level may vary systematically through most of the cleavage stages within an area of less than a square mile. High levels of strain are located along the hinges and east limbs of major folds in the central part of the Blandford-Woronoco area and in the synform in the south-eastern part of the area (E on fig. 3). The strain level appears to drop off both to the north and west. The major antiform in the central part of the Blandford-Woronoco area was in part developed during F4 time (B on fig. 3). The high level of strain represented by F3 folds on its eastern limb is thus a product of both the strain associated with F3 and F4.

1/ Designations for cleavages which are paralled to the axial surfaces of minor folds include the terms schistosity, crenulate cleavage, slip cleavage, and spaced schistosity. The term schistosity is restricted to cleavages where the micas are all essentially oriented parallel to the cleavage plane in the rock. Crenulate cleavage is formed by the parallelism of short limbs of crenulate folds in schist. Although discrete planes of slip or fabric discontinuity are not present on the scale of the thin section or hand specimen, a distinct surface where the micas change orientation abruptly does exist on the scale of the outcrop. Slip cleavage is a more advanced stage of crenulate cleavage. Here discrete slip has occurred across the cleavage plane and the rock commonly breaks along this surface. In both crenulate cleavage and slip cleavage the micas are oriented at an angle to the cleavage plane. Spaced schistosity refers to an advanced stage of slip cleavage in which the rock is divided into two fabric domains - the actual cleavage planes where the micas are parallel or closely parallel to the cleavage, and the intercleavage domains where the micas are discordant to the cleavage. Here the micas form the remains of small crenulate folds. Crenulate cleavage, slip cleavage, and spaced schistosity are part of an overall sequence of cleavage development. Each stage represents a higher level of strain. The boundaries between these categories are gradational and quite arbitrary since they have not been quantified. Although this sequence is commonly typical of F3, there is sufficient evidence to indicate that it is the dominant mechanism of cleavage development in all four fold generations.

F4 folds represent a relatively low level of strain with a restricted range. Generally, the minor folds are broad undulations that only develop into fairly tight crenulate folds along the antiform outlined by the eastern bend in the base of the Goshen Formation (symbol DSg, fig. 5).

Axial surface maps

Axial surface maps have been prepared for the youngest three generations of folds in the Blandford-Woronoco area, no such map for F1 was prepared because the number of reliable observations is few. However, F1 folds are numerous in the lower member of the Cobble Mountain Formation, since well-preserved graded beds are either normal or inverted on minor antiformal folds of F2.

As shown on the axial surface maps, most of the minor fold data are confined to the Ordovician rocks in the Blandford-Woronoco area, because the outdrop is very abundant in this area and is more limited to the west. The area underlain by Silurian and Devonian rocks has been mapped by Hatch and S. F. Clark, Jr. and is excluded from this study. My work, however, has included a detailed study of the basal 100 meters of the Silurian and Devonian section.

F2 folds:

Figure 1 is based on 270 axial surfaces of F2 minor folds distributed rather evenly in the area of study. The axial surfaces of these minor folds are systematically folded by most of the major folds labelled by letters A through F in figure 1 and, hence, are older than folds A-F. Two of these younger folds (A and D) are outlined by the Taconic unconformity. In the pre-Silurian rocks a large refolded antiform outlined by the contact between the lower and upper members of the Cobble Mountain Formation is the largest and most conspicuous major structure of F2 age. The nose of this fold, here called the Woronoco fold, is situated at locality 1 (fig. 1) where the schistosity trends at right angles to the bedding, and almost totally obliterates it. A smaller F2 hinge is mapped at locality 2 (fig. 1). Close inspection of the contact between the two members of the Cobble Mountain Formation (Ocl and Ocu, fig. 1) shows that south of locality 1 the axial schistosity of F2 folds cuts the map contact (bedding) at an acute angle with a sinistral sense (as indicated by an arrow drawn from the axial surface normal to the bedding normal in spherical projection). North of locality 1, except just south of locality 2, the sense has changed to dextral. The generalized trace of the axial surface of this major antiform (heavy line on fig. 1) is located in the lower member of

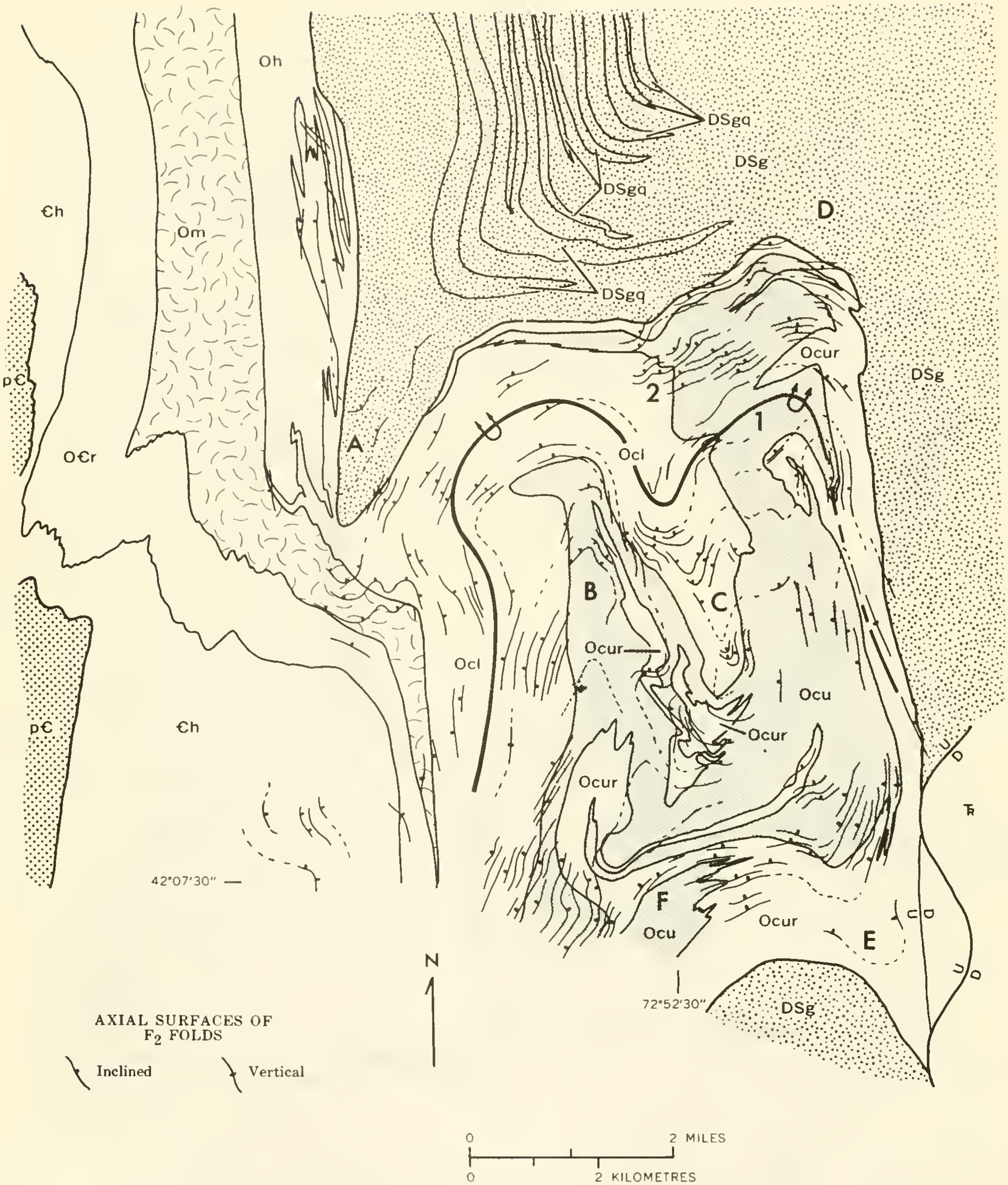

Figure 1. Geologic map of the Blandford-Woronoco area showing axial surfaces of F2 folds. Stratigraphic units and symbols are listed in Stanley 1975; also, pꞒ, Precambrian rocks; Ocur, rusty schist and gneiss in the upper part of the Cobble Mountain Formation. Letters A through F and numbers 1 and 2 are discussed in text. Heavy curved line marks axial surface of Woronoco fold, outlined by the contact between Ocl and Ocu.

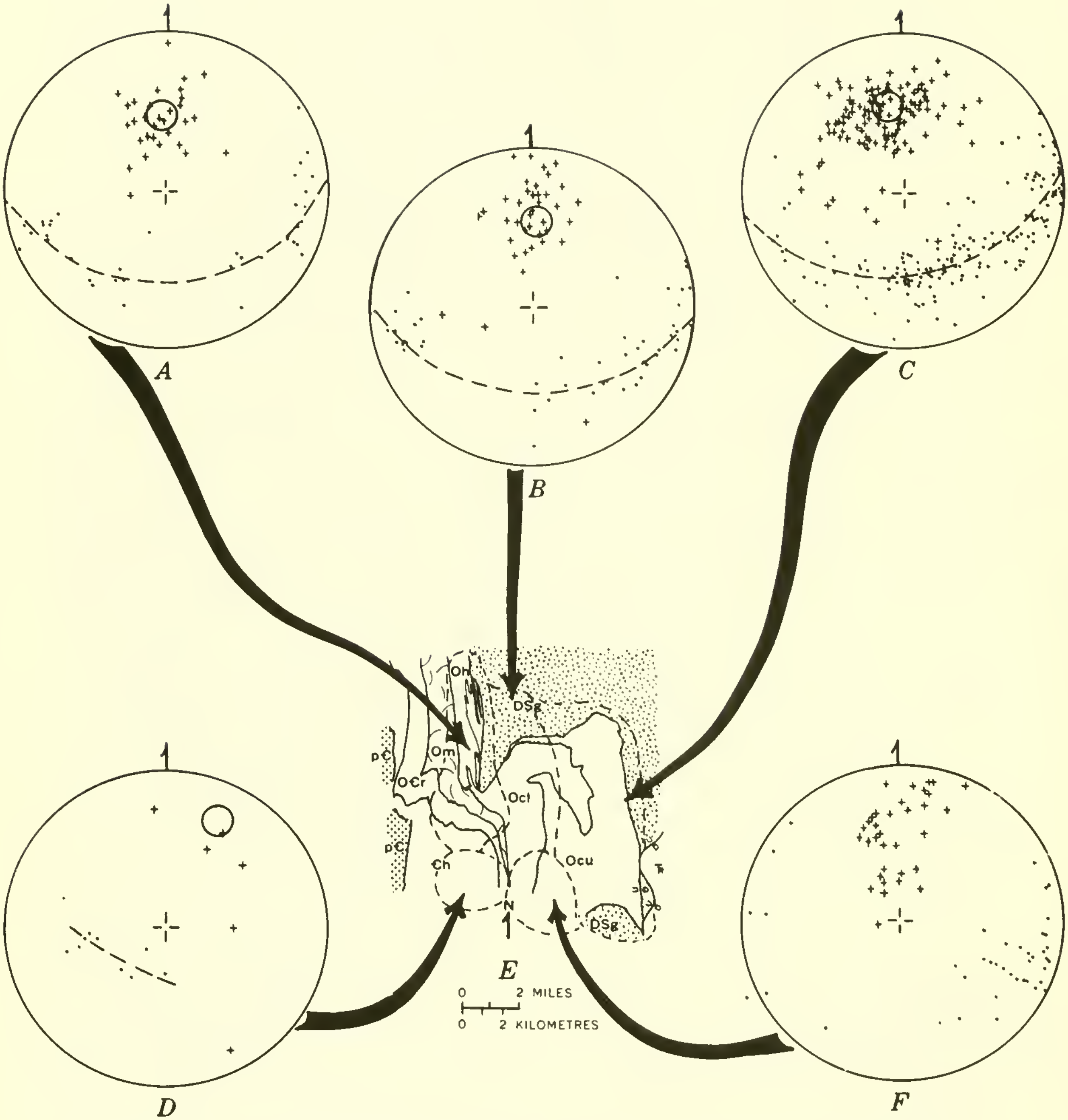

Figure 2. Lower hemisphere equal-area projections of poles to F2 axial surfaces (·) and F2 fold axes (+) for 5 subareas in the Blandford-Woronoco area. The dashed lines are great circles that approximate the distribution of poles to axial surfaces. The pole to the great circle (π pole) is shown by the center of the 1 percent circle. Subarea A, 40 axial surfaces, 26 fold axes; subarea B, 47 axial surfaces, 23 fold axes; subarea C, 165 axial surfaces, 107 fold axes; subarea D, 12 axial surfaces, 6 fold axes; subarea F, 42 axial surfaces, 32 fold axes. Plane of projection is horizontal with north indicated by arrow.

the Cobble Mountain Formation and is systematically folded by the younger antiform B and synform C (fig. 1). The nose of this older major fold along the contact at the base of the Silurian and Devonian section appears to be somewhere to the south under the Triassic rocks, because the axial schistosity cuts this contact at an acute angle with a dextral sense.

The axial-surface schistosity of F2 folds is identical to the regional schistosity that parallels the axial surfaces of the isoclinal folds in the Goshen Formation described by Hatch (1968, 1975). This relationship is best seen along the eastern side of the Blandford-Woronoco area south of the letter D on figure 1. The minor folds of F2 are, therefore, coeval with the folds of stage 2 described by Hatch. Although the unconformity at the base of the Goshen Formation in and north of the Blandford-Woronoco area is apparently undeformed by the major isoclinal folds that fold the Goshen, the base of the Silurian and Devonian section to the south in western Connecticut is highly folded around and in between the gneiss domes (Stanley, 1975).

Many of these folds appear to be correlative with F2 in the Blandford-Woronoco area because the regional schistosity is parallel to their axial surfaces and the axial surfaces of isoclinal minor folds delineated by graded beds in the Silurian and Devonian rocks. The axial surface of one of these major folds is shown in the southern part of the Blandford-Woronoco area in figure 1.

Fold axes and poles to axial surfaces of F2 folds are shown for five subareas in figure 3. In diagrams A, B, and C, the π pole for each axial surface girdle is oriented within the cluster of points representing fold axes. This relationship indicates that the axial schistosity of F2 is folded about younger axes that are generally parallel with the fold axes of F2. Local departures from parallelism, however, do exist at the outcrop. Because the π poles for subareas A, B, and C are nearly parallel, the axes of subsequent folding was essentially parallel throughout most of the area in figure 1. Diagrams D and F are small subareas where the π pole is either based on insufficient data (D, fig. 2) or appears to be steeply inclined (F, fig. 2).

F3 folds:

Figure 3 is based on the axial surfaces of 900 F3 folds of which 650 are plotted as data points for the axial surface map. Areas of high strain level are located in the major folds at H, in the fold at F, on the east limb of antiform B, and in the synform which is continuous from C to E (fig. 3).

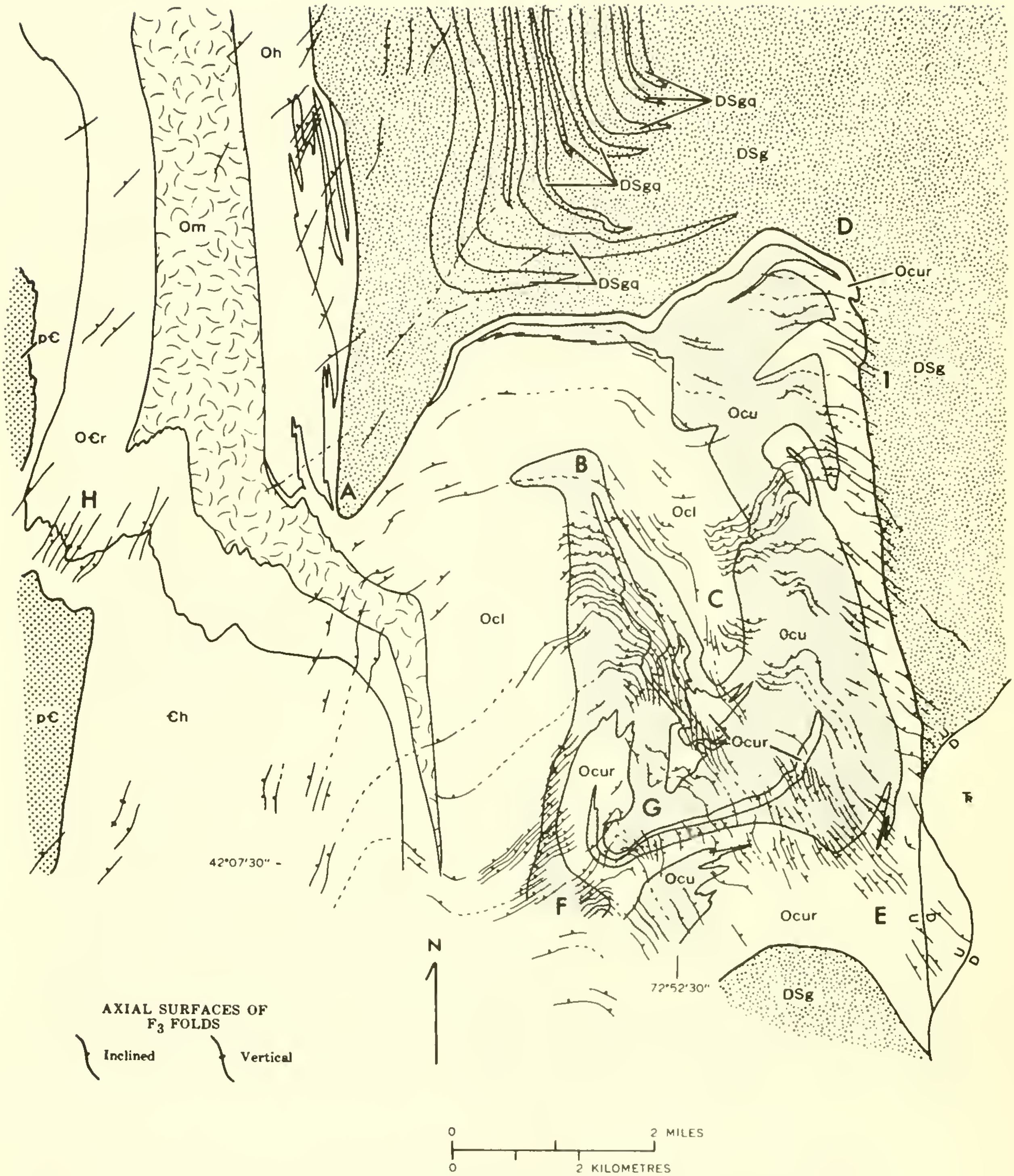

Figure 3. Geologic map of the Blandford-Woronoco area showing axial surfaces of F3 folds. Stratigraphic units and symbols are listed in Stanley (1975) and described in figure 1. Letters A through H and number 1 are discussed in text.

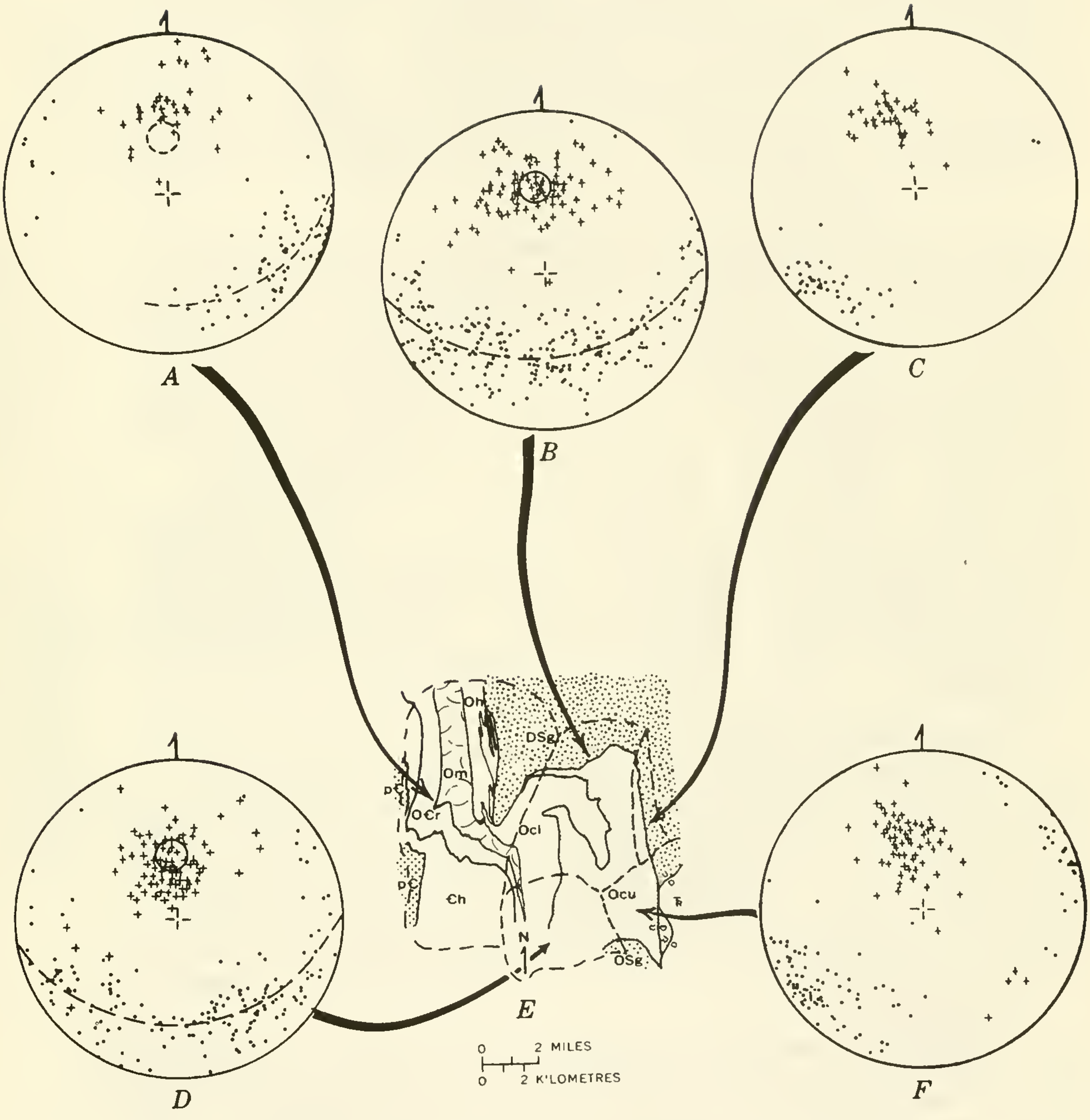

Figure 4. Lower hemisphere equal-area projections of poles to F3 axial surfaces (•) and F3 fold axes (+) for 5 subareas in the Blandford-Woronoco area. The dashed lines are great circles that approximate the distribution of axial surface normals. The center of the 1 percent circle (π pole) is the pole to the great circle. Subarea A, 79 axial surfaces, 32 fold axes; subarea B, 235 axial surfaces, 85 fold axes; subarea C, 46 axial surfaces, 32 fold axes; subarea D, 158 axial surfaces, 83 fold axes; subarea F, 118 axial surfaces, 68 fold axes. Plane of projection is horizontal with north indicated by arrow.

In these areas the slip cleavage is well developed and the folds are tight. A spaced schistosity is developed at F (stop 7) and some of the other localities. The axial surface cleavages of F2 and F3 are nearly parallel along the east limb of antiform B, but can be separated here by their differences in form. Perhaps the greatest divergence in orientation is in synform E, where the axial surface of F3 trends northwestward across the folded axial surfaces of the older F2 folds (compare figs. 1 and 3).

The axial surfaces of F3 folds are systematically deformed into a series of broad cleavage antiforms and synforms. These folds become a single broad antiform or arch to the north near the contact between the Silurian and Devonian rocks and the underlying formations (north of B on fig. 3). The large folds at B and D and the smaller fold just northeast of C are clearly younger in age than F3 since they fold this cleavage. The major fold at A, the folds to the south and west of A, the folds at H, part of the fold at F, the major synform that is continuous from C to E, and the small fold at 1, are all coeval with F3 minor folds, because the calculated axial surface of these mappable structures is statistically parallel to F3 minor folds. The two major folds (A and D) along the Taconic unconformity are, therefore, of two quite different ages, although their general orientation seems identical on the map.

Fold axes and poles to axial surfaces of F3 folds are shown for five subareas in figure 4. These subareas do not coincide with those for F2 folds, although considerable overlap does exist. In three of the five subareas the π poles to the axial surface girdles fall within the cluster of fold axes, thus indicating that the axes of F4 major folds are generally parallel with the axes of the older F3 folds. It thus appears that on a subarea basis the π poles of F2 and F3 axial-surface girdles are generally parallel or nearly so.

F4 folds:

Figure 5 shows axial surfaces of 106 minor F4 folds, of which 32 deform the slip cleavage of F3 folds (stop 7, fig. 9). The remaining 74 crenulate folds are assigned to this generation because their axial surfaces are parallel to the 32 known F4 folds and discordant to F3 folds, although the two sets are not directly superposed.

F4 folds which directly fold the older slip cleavage (heavy barbell symbol, fig. 5) are best developed and more numerous along the axial traces of major F4 folds (compare figs. 3 and 5). The strain associated with F4 folds is slightlymore intense along the easternmost antiform than those to the west.

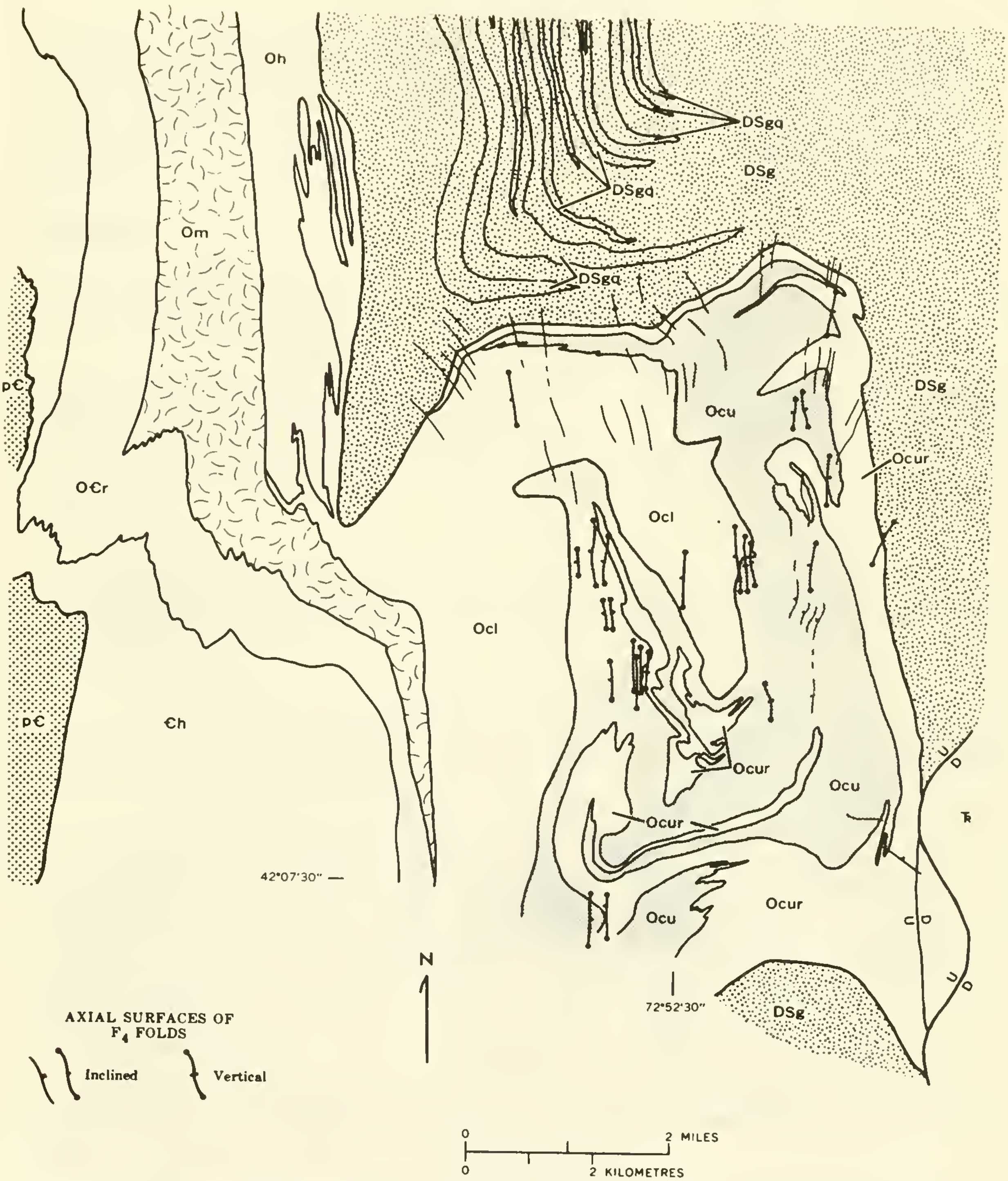

Figure 5. Geologic map of the Blandford-Woronoco area showing axial surfaces of F4 folds. Heavy barbell symbols represent F4 folds that deform slip cleavage of F3 folds. Light symbols are F4 folds that deform F2 schistosity. Stratigraphic units and symbols are listed in Stanley (1975) and described in figure 1.

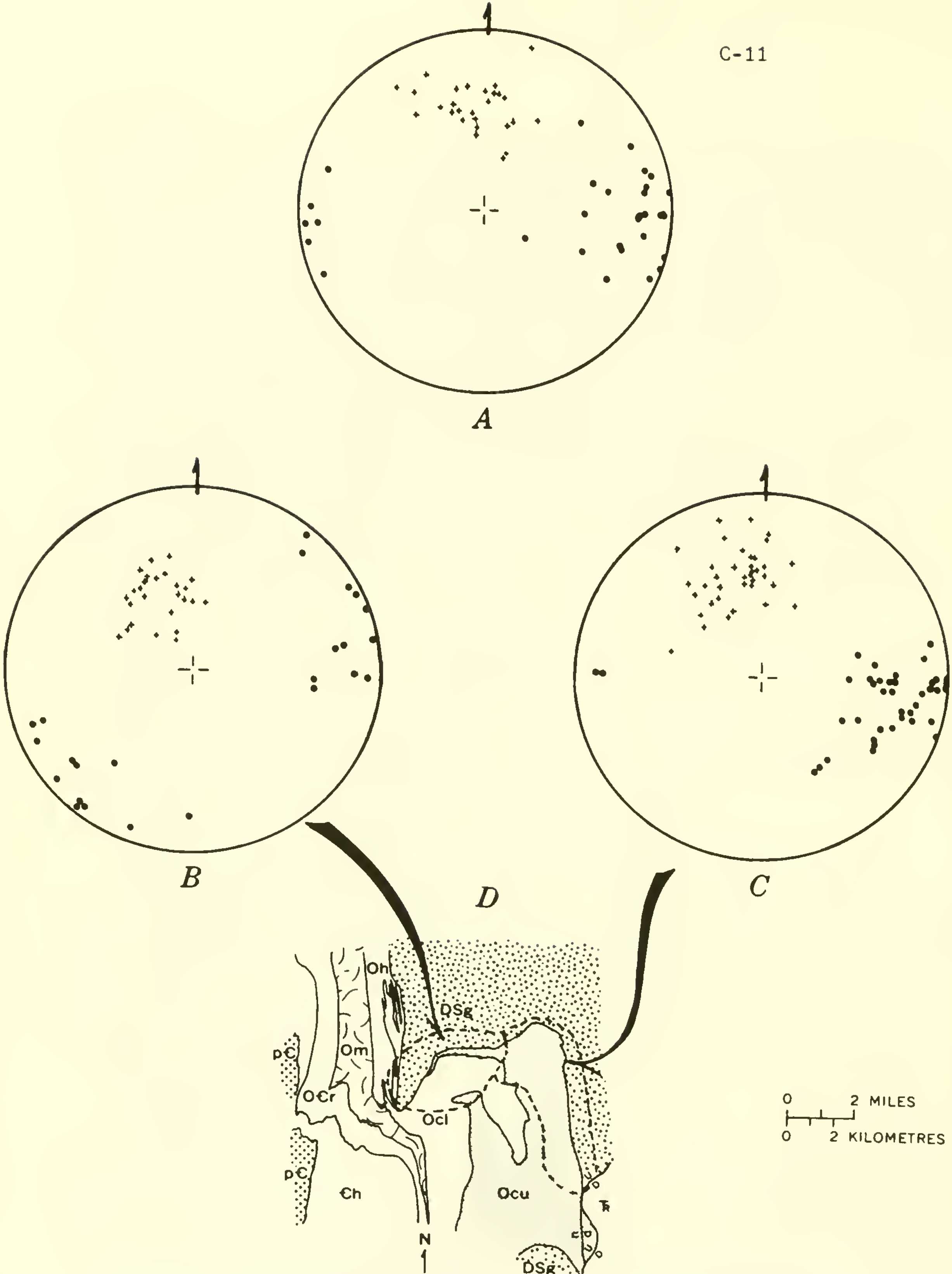

Figure 6. Lower hemisphere equal-area projections of poles to F4 axial surfaces (•) and F4 fold axes (+) for 2 subareas in the Blandford-Woronoco area. Diagram A contains folds that deform the slip cleavage of F3 folds in both of the areas shown by B and C. Diagrams B and C contain folds that deform only F2 schistosity. Subarea A, 31 axial surfaces, 27 fold axes; subarea B, 27 axial surfaces, 26 fold axes; subarea C, 41 axial surfaces, 37 fold axes. Plan of projection is horizontal with north indicated by arrow.

The axial surfaces of F4 folds directly superposed on F3 folds trend to the north and dip steeply to the west and are very similar to the crenulate folds in subarea C along the eastern part of the area. To the west, however, the axial surfaces trend more westerly. The F4 axes are remarkably similar in orientation to those of the older fold generations.

Summary

The previous discussion of fold generations has suggested the following important conclusions:

1. Axial surfaces of minor folds provide a means of determining the relative age of major structures which otherwise may be ambiguous or indeterminate. If the axial surfaces of a given fold generation are statistically parallel to the calculated axial surface of the major fold, then the two are coeval. Older and younger ages are based on cross-cutting relations.

2. The angular relationship between formation contacts and the axial surfaces of minor folds of a given age can locate the hinge and limb areas of refolded major folds of the same generation.

3. The character or style of axial surface cleavage (i.e., crenulate cleavage, slip cleavage, spaced schistosity) in a specific rock type is a reliable but qualitative measure of strain for each fold generation.

4. The axes of minor folds and major cleavage folds for F2, F3, and F4 are essentially parallel throughout most of the Blandford-Woronoco area. Therefore, deformation has involved repeated folding about parallel axes with the development of discordant axial surface cleavages similar to TYPE 3 interference pattern of Ramsay (1967, p. 531-535). This constraint simplifies subsequent kinematic and dynamic analysis.

5. The base of the Silurian and Devonian section marks a surface of structural décollement inasmuch as the ratio of amplitude to wave length is much higher in the younger rocks than in the pre-Silurian section.

Structural Evolution and Plate Tectonic Speculations

Any model of the structural evolution for the Blandford-Woronoco area must be highly qualitative and speculative. However, an internally consistent and reasonable geometric sequence based on the data presented in the foregoing pages is diagrammatically shown in the profile section in figure 7. The control for diagram 4 in figure 7 is quite good, although

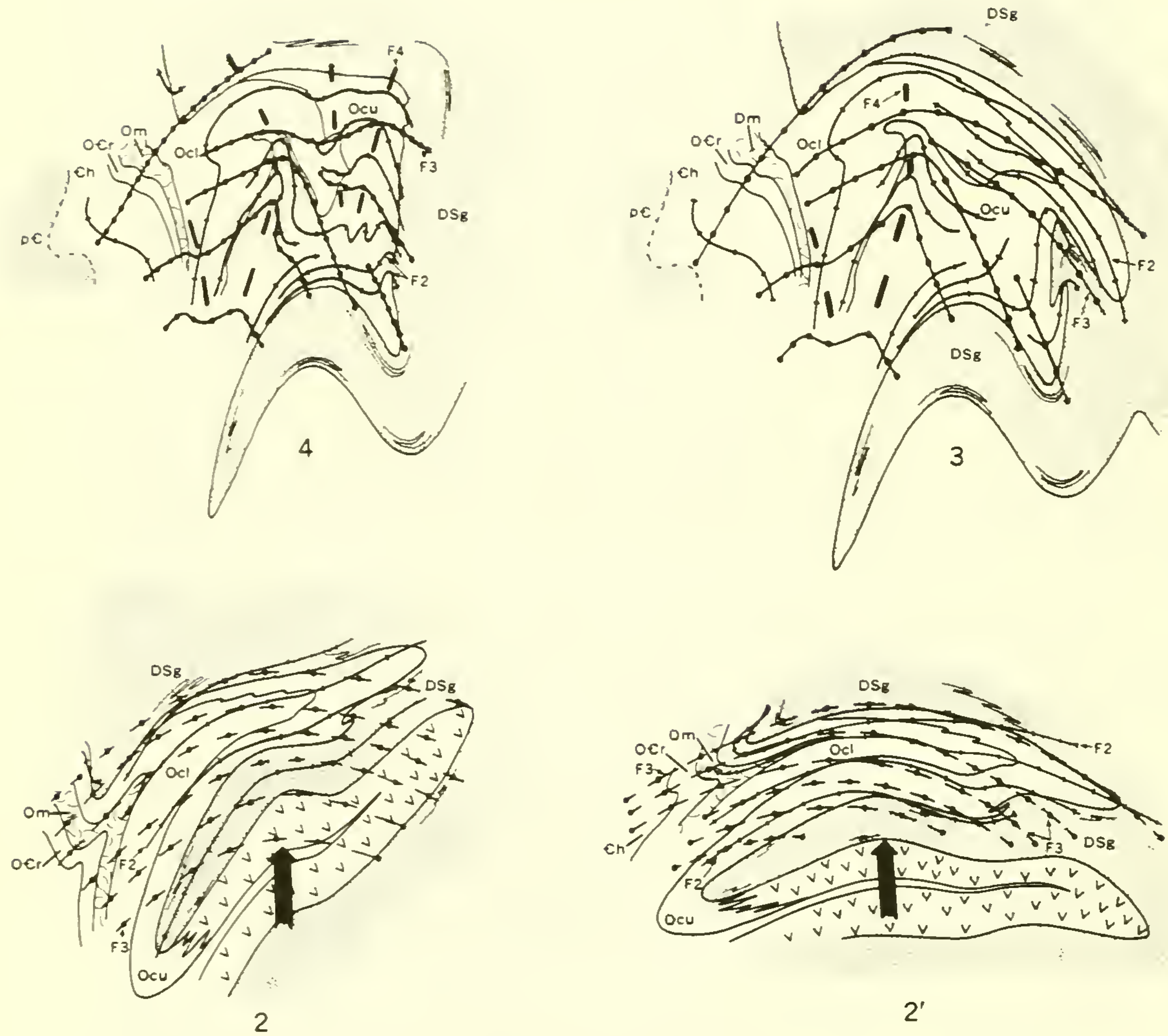

Figure 7. Structural evolution of the Blandford-Woronoco area with emphasis on generations 3 and 4. Diagrams are profile sections drawn perpendicular to an axis oriented approximately N5W at 45 degrees. Axial surfaces of each generation identified by the symbols F2, F3, and F4. Diagram 4 shows reconstruction prior to Triassic faulting; diagrams 4, 3, 2, (2') show progressively older configurations; diagrams 2 and 2' are alternative interpretations of early F3 time. Granville dome is located approximately in the position of each diagram number.

the effects of Triassic faulting have been removed. The remaining diagrams are derived by simply externally rotating the limbs of F4 folds about F4 axes until the axial surfaces of F3 are essentially planar. F2 structures are not unfolded in figure 7.

The configuration in diagram 4 represents a profile section of the Blandford-Woronoco area at the close of F4 time. The axial trace of the largest F4 cleavage antiform is offset to the west of the symmetry plane of the Granville dome (located directly over the number in each diagram).

Diagram 3 represents a stage in the progressive unfolding of F3 axial surfaces. Here the eastern limb is straightened out, forming one large cleavage anticline. In so doing the easternmost contact between pre-Silurian and younger rocks is folded into a large inverted anticline which is required by the Woronoco fold of F2 age (fig. 1). By continuing to unfold the axial surfaces of F3, the cleavage antiform of diagram 3 is reduced to the very broad arch of diagrams 2 and 2'. This arch is essentially parallel to the direction of maximum extension developed in mantling rocks as a result of the emplacement of a mantled gneiss dome, according to the mathematical model of Fletcher (1972, p. 210). During this early phase in F3 time, felsic volcanic rocks of Middle Ordovician age are thought to have moved upward, forming elongate domes and modifying the older isoclinal folds. Evidence supporting this event is found to the south in western Connecticut where these volcanic rocks appear to form the cores of several large east-facing folds that have been redeformed into gneiss domes outline by the contact between the Cambrian-Ordovician and the Silurian-Devonian rocks of figures 1 and 16 (Stanley, 1975; Dieterich, 1968a and b). In the Blandford-Woronoco area F3 folds are considered to be a product of the emplacement of the granville dome, because their axial surfaces form a cleavage antiform or arch that partly bridges the dome. Analogous structures are found in eastern Vermont where a late crenulate-slip cleavage forms the distinct arch of the Strafford-Willoughby arch (White, 1949, figs. 2 and 3). F3 folds are particularly well developed in synclines between the domes in western Connecticut (Stanley, 1975). These synclines probably developed when the domal rocks moved upward causing the intervening areas to be squeezed together.

Diagrams 2 and 2' are alternative interpretations on the attitude of the older isoclinal folds during early F3 time. In both diagrams the upward movement of the lighter volcanic rocks resulted in a fan-shaped compression field in the overlying rocks. Diagram 2, with moderate west-dipping axial surfaces, is preferred since the axial surfaces of F3 intersect rock units at a fairly high angle just north of the Granville dome (fig. 3). This is well illustrated

in the lower parts of diagrams 4 and 3 just above the refolded syncline of Silurian and Devonian rocks. In diagram 2' the axial surfaces of F2 and F3 intersect at a very small angle. Furthermore, F3 folds with their associated cleavage are far more likely to form if the schistosity of F2 is compressed at a low angle to the schistosity rather than at a high angle. In diagram 2' the older schistosity would be compressed at a high angle and hence the tendency to fold it would not be as great.

Older configurations can be visualized by reducing F3 axial surfaces to planes. Although a separate diagram was not prepared for F1, the appropriate form can be imagined by unfolding the isoclinal folds and removing the Silurian and Devonian section. The development of folds in the Blandford-Woronoco area can then be visualized by following the sequence of diagrams from 2 to 4 in figure 7.

In summary, the structural evolution of the Blandford-Woronoco area in Phanerozoic time began with development of F1 folds in rocks of Middle Ordovician and older age. Although structures older than F2 and younger than F1 have been reported by Osberg (1972, 1975) in the Silurian and Devonian rocks, they do not appear to be the same age as the F1 folds. It is tentatively concluded, therefore, that F1 folds developed during the Taconic orogeny. Few if any large-scale folds of this generation have been recognized in western Massachusetts, although some small ones have been reported (Hatch, 1975) and several small folds in the upper member of the Cobble Mountain Formation are believed to be of this age (fig. 1). F1 folds may well have resulted from intense shearing due to westward thrusting of Cambrian and Ordovician rocks along the root zone of the high Taconic thrust during Middle or Late Ordovician time (Norton, 1971). This root zone appears to continue southward into western Connecticut where it is inpart represented by Cameron's Line (Hatch and Stanley 1974, Plate 1).

The remaining three generations of folds are well developed in the Silurian and Devonian section and are a result of Acadian deformation, since Acadian metamorphism outlasted two if not three of the fold generations (Hatch, 1975). This deformation has completely remolded and largely obscured the older structures. The isoclinal folds of F2 have resulted from severe east-west compression in which the axis of maximum compression was inclined at a gentle to moderate eastward angle. The intense transverse compression is compatible with active plate collision during the Middle Devonian. The westward inclination of the axial surfaces and consequent east-facing sense of the isoclinal folds may well result from local underthrusting of the orogen beneath the eastern edge of the continent.

The subsequent rise of the light felsic rocks in the Middle Ordovician section produced the gneiss domes which in turn refolded the older isoclinal folds of F2 and may reflect a pause in plate interaction. East-west compression during F4 time formed large-scale cleavage folds. Although F4 folds are fairly extensive regionally, they are only well developed locally and, hence, may reflect the irregularities of plate boundaries. In Late Triassic time normal faulting associated with the opening of the Atlantic cut the eastern part of the area.

As a closing note, I recommend the following poem:

Synclines

Synclines become synclines
  Through tens of millions of years
Of downwarping, downthrusting
  and gentle sagging.
Often to be folded
And finally eroded,
Leaving no tombstone.

A faded relict was brought to light
  By circumstantial chance.
I dined upon it, eating my bologna sandwich
And marked its outcrop upon my map.

While surmising my morning's work
  And daydreaming into nowhere,
A ray of sunlight cascaded
  Onto my old gray outcrop
And framed a lineation.

It was indeed a syncline,
  One I'd missed to see
Through metamorphism
  Of deep woods mystery.

It made my vision clearer
  Of what had happened here
In Ordovician time
  Back four hundred fifty million years.

The syncline had been folded,
  then refolded twice again
Each fold impressed upon the other
  With axial cleavage S_1, S_2, S_3.

So much it told me,
  Out there in the field
That I never expected more in the lab
  Than a basic minerologic yield.

But two months hence
Under 40X
A glaucophane appeared
  Cloaked in scientific suspense.
Then another, and one or two more
All associated, it seemed, with fold generation two.

So much it told me
That single syncline,
After waiting four hundred fifty million years.
Like an old man on his deathbed
  Reciting ancient tales
Of Custer's Last Stand
  And Abraham Lincoln
All mottled and confused.

Just a simple syncline
  Highly metamorphosed
That will weather into history
  In another million years.

Robert Badger, 1975
University of Vermont

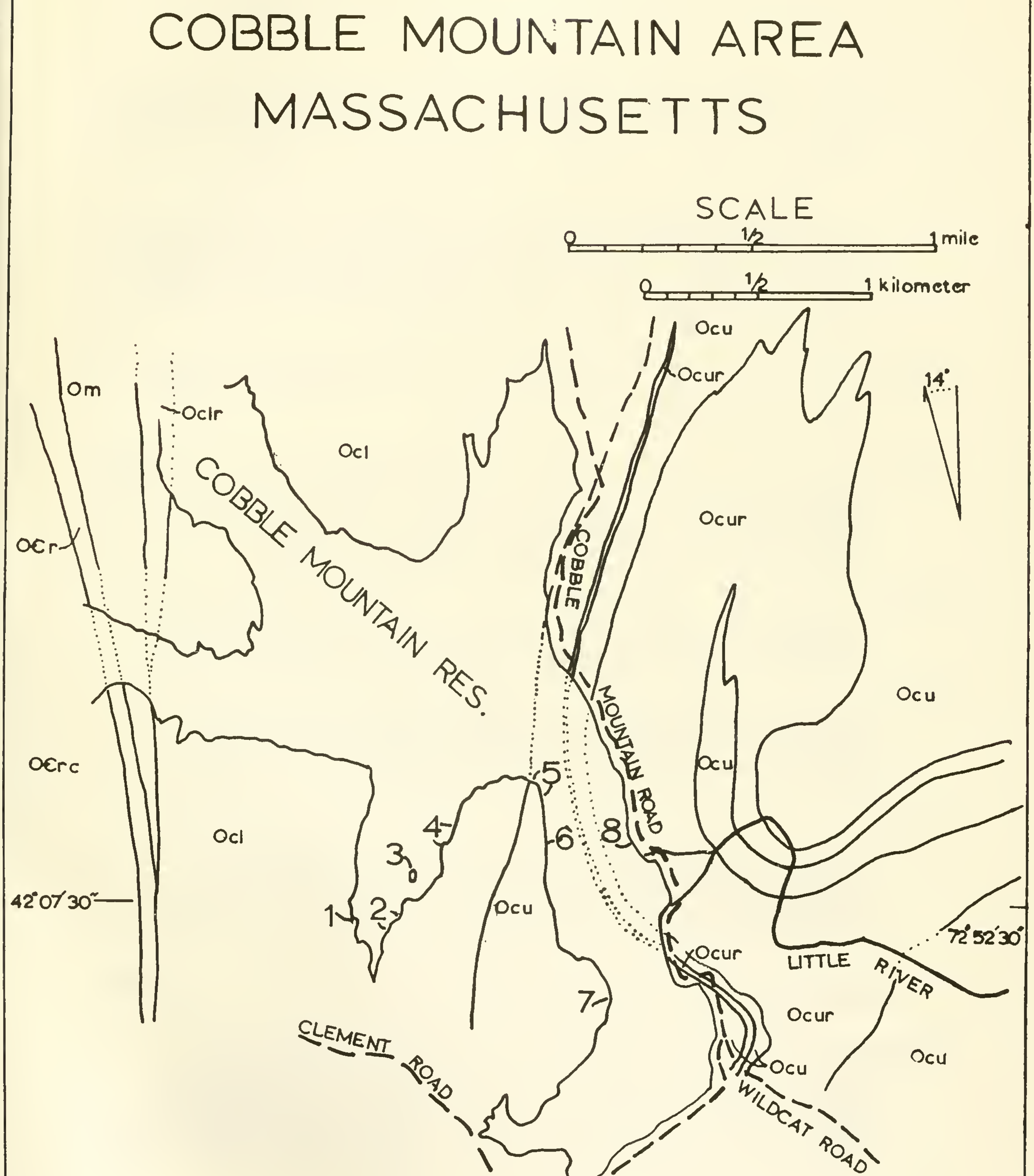

Figure 8. Traverse map for Stops 1-8.

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